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GUIDE TO THE GRANITES
AND MINERAL DEPOSITS
OF SOUTHWESTERN NOVA SCOTIA

Editors

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PREFACE

This guidebook is the first draft of Nova Scotia Department of Mines and Energy Paper 85-3. It is being issued in this preliminary form, before completion of the editorial work, in order to complement the field trip to southwestern Nova Scotia associated with the CIM Conference on Granite-Related Mineral Deposits. The Editors would appreciate comments towards improving the final version. Had it not been for J. Bates, K. Mills, R. Jones, B. MacDonald, S. Saunders, M. Wolfe, J. Cumby, D. Bernasconi, J. Campbell, P. Belliveau, C. Phillips, C. Murphy, W. Burt, G. O'Reilly and P. Giles, this undertaking would not have been completed.

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CHAPTER 1

THE SOUTH MOUNTAIN BATHOLITH: GEOLOGY, PETROLOGY, GEOCHEMISTRY

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ABSTRACT

The South Mountain Batholith (SMB) outcrops over 10,000 km² in the Meguma Zone and is the largest peraluminous granitoid complex in the Appalachians. It has a three-dimensional mushroom shape, and is a typical massive, post-tectonic epizonal granitoid characterized by a discordant relationship with the country rocks, a pronounced thermal metamorphic aureole and microplitic cavities. Petrologically, it consists largely of biotite granodiorite, two-mica monzogranite and leucomonzogranite showing a variety of textural types, and contains a suite of minerals characteristic of peraluminous granitoids (andalusite, garnet, cordierite, tourmaline, topaz, etc.). Geochemically, major and trace elements show that the entire suite is comagmatic, with the exception of small areas subsequently affected by strong metasomatic alteration (paraintrusive suites) which are often associated with economic mineral deposits. Evidence from both radiogenic and stable isotope geochemistry points to a primitive crustal, but not a Meguma flyschoid metasedimentary, source for the magmas. The possibility of a mantle component in the SMB cannot yet be excluded.

REGIONAL SETTING

The SMB is the largest (>10 000 km²) peraluminous granitoid complex in the Appalachians, and it outcrops in the Meguma Zone of southern Nova Scotia (Fig. 1-1). Over most of the area it is in contact with Cambro-Ordovician siliclastic metasediments of the Meguma Group which had undergone lower greenschist to lower amphibolite facies metamorphism prior to its intrusion. The cumulative stratigraphic thickness of the Meguma Group has been estimated to be in excess of 15 km (Schenk, 1971, 1978) and may be substantially more because of structural thickening. Nowhere is the base of the group exposed. Along the northwestern margin younger sediments ranging up to Lower Devonian (Emsian) are intruded. Elsewhere Tournaisian sediments nonconformably overlie the SMB.

Although the long axis of the complex approximately parallels the trend of the Acadian regional structures, on a local scale contacts are clearly discordant and, together with textural evidence from the contact aureole, point to a post-tectonic emplacement.

FIELD RELATIONSHIPS

All phases of the SMB are massive and generally unfoliated. Where aligned textures are evident they have been attributed to flowage phenomena or crystal settling (Clarke and Muecke, 1980; Abbott, 1982). When contact relationships with the country rocks can be studied, no disturbance of regional trends is evident, but there is ample evidence of stoping. Meguma-like enclaves are particularly abundant in the granodiorites and their abundance shows no clear correlation with proximity to the country contact. In the more leucocratic rocks enclaves are distinctly rare.

A very large segment of Meguma metasediments (5 km by 20 km) in the centre of the SMB, near the New Ross-Vaughan Complex (NRVC), shows structural trends which conform to the regional trend. This possible roof pendant, enclave abundance, microplitic cavities, discordant contacts and an extensive thermal aureole all suggest epizonal intrusion at a depth in the range of 1.5 to 3 km during which brittle processes in the form of stoping predominated as the intrusive mechanism.

The internal structure of the batholith is characterized by a plethora of petrologically distinct phases, with bodies ranging in size from batholiths to thin dykes. Biotite granodiorite is the most voluminous phase in the SMB and is intruded by a number of smaller bodies of biotite-muscovite monzogranite (adamellite), porphyritic monzogranite (porphyry), leucocratic monzogranite (leucadamellite), as well as minor bodies of aplite and pegmatite (McKenzie and Clarke, 1975; Muecke and Clarke, 1981). The latter two phases usually show sharp intrusive contacts whereas contacts among the others are rarely observed and are apparently more gradational. Where the relationships are unambiguous the more mafic phase is always cut by more leucocratic phases.

GEOPHYSICAL DATA

Gravity surveys by Garland (1953) detected strong negative Bouguer anomalies (up to -55 mGal) in the eastern part of the SMB and suggested that the SMB has a mushroom shape averaging about five km in thickness along the margins and attaining local thicknesses in the order of 25 km (eg. the NRVC). More recent three-dimensional models (O'Reilly, 1975; Keen, 1976) essentially confirm such a structure, but refine the dimensions to 10 and 20 km respectively. Expansion of the surveys over the western batholith has revealed a similar, but less extensive, anomaly centered over the Davis Lake pluton (Darnley, 1982).

Airborne radiometric surveys reveal distinct anomalies for eU, eTh, eK and eU/eTh. The most pronounced of these are centered over the NRVC and the Davis Lake pluton (Darnley, 1982; Chatterjee and Muecke, 1982). High heat flow values have been obtained over the NRVC (Hyndman et al., 1979) and suggest that the high radioelement concentrations may extend to considerable depth. Chatterjee and Muecke (1982) pointed out that areas with high total radioactivity and high eTh values within the batholith correlate with granite bodies showing major tin-uranium mineralization.

Magnetic surveys over the Meguma Terrane show a very strong contrast between the often highly magnetic metasediments of the Meguma Group and the very weakly magnetic granitoids. The scarcity of primary ilmenite and absence of magnetite in the granitoids conform with these observations.

PETROLOGY

The most mafic and most abundant lithology in the better known eastern part of the SMB is a biotite granodiorite which outcrops over ~75% of the area

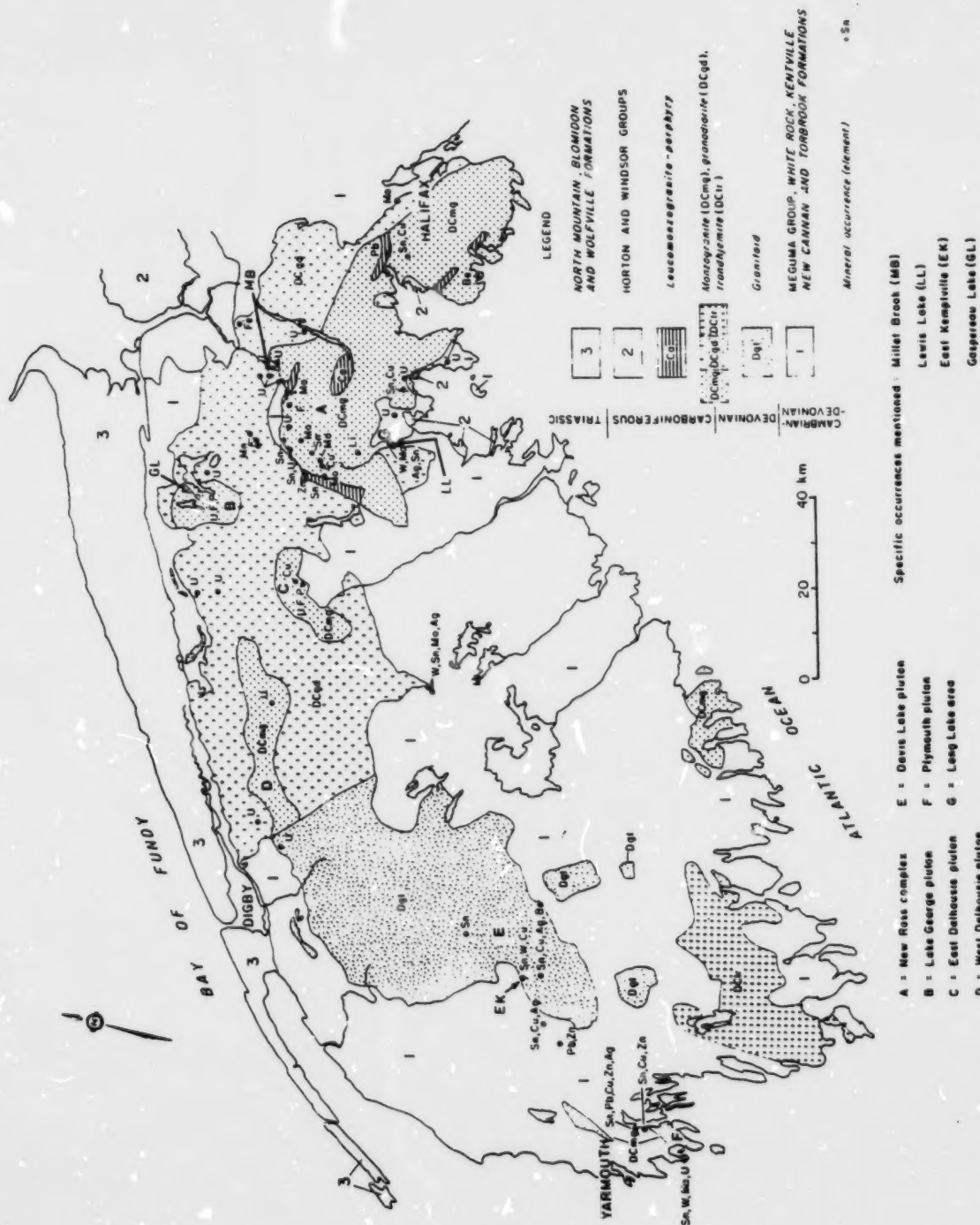


Figure 1-1. Geological map of southwestern Nova Scotia showing the major granitoid lithologies in the South Mountain Batholith and the locations of known mineral deposits (after Chatterjee and Kuecke, 1982).

(McKenzie and Clarke, 1975). It is characterized by the presence of a single primary mica, biotite and the dominance of andesine. Biotite constitutes on the average 12.5% of the mode in a coarse-grained hypidiomorphic-granular texture.

Monzogranites and porphyries are second in areal extent and are intruded into the granodiorites. They are recognized by the presence of two micas, muscovite and biotite, which together generally account for <10% of the mode. Potassic feldspar predominates and andalusite is a common accessory phase. A large textural diversity is evident in this group and ranges from porphyritic aplites ('porphyry') to coarse grained porphyritic or non-porphyritic rocks.

The most leucocratic members of the complex occur as dykes and minor intrusive bodies which are mainly associated with monzogranitic rocks with which they are in sharp intrusive contact. The suite includes aplite-pegmatite dykes and pods, leucomonzogranite bodies, and highly differentiated porphyries. They are texturally diverse, can have complex mineralogies, and are frequently accompanied by extensive hydrothermal and metasomatic interaction with the host.

A metasomatic ('paraintrusive') suite (Chatterjee and Muecke, 1982) is usually spatially and probably genetically associated with the monzogranites and leucomonzogranites. Albitized granites, albitites, episyenites, argillized and sericitized granites, biotite leucogranites, and various types of greisen are the common members of this group. Extensive interaction of fluids with residual magma or crystalline rocks or both are thought to account for such lithologies (Chatterjee and Muecke, 1982; Farley, 1979; Logothetis, 1984).

MINERALOGY

By definition quartz, alkali feldspar and sodic plagioclase are essential constituents of members of the granitoid family regardless of chemical affiliation or origin. As the whole-rock ratio of A/CNK (mol $Al_2O_3/(CaO+Na_2O+K_2O)$) varies from less than one (peralkaline) to greater than one (peraluminous), a distinctive assemblage of accessory minerals, called the characteristic minerals, occurs (Clarke, 1981). In the case of the peraluminous granites, these characteristic minerals can include biotite, muscovite, garnet, cordierite, aluminosilicates (particularly andalusite and sillimanite), tourmaline, topaz, mullite, spinel and corundum, all of which can nucleate directly from a granitic melt, but may have other origins also. Many of these characteristic phases occur in the SMB and are discussed below. Other rare accessory phases such as triplite (Mandarino et al., 1984), ilmenite, columbite-tantalite, and niobian rutiles are also mentioned briefly.

BIOTITE

The modal abundance of biotites decreases and $FeO+Fe_2O_3/(FeO+Fe_2O_3+MgO)$ ratio increases from the early granodiorites (10-20%; 0.78-0.84, respectively), through the monzogranites (5-12%;

0.80-0.90), to late aplites and porphyries (0-4%; > 0.90) (Stallard, 1975). All of these biotites are apparently of magmatic origin, although the very high concentrations in some of the granodiorites may suggest either concentration during crystal-liquid fractionation, or incorporation of xenocrystic material. If the latter, then the more magnesian biotites of the aureole rocks have become completely re-equilibrated in the granitic melt prior to solidification. In the hydrothermally altered rocks, primary biotite commonly breaks down to various combinations of chlorite, phengite, K-feldspar and rutile. A new generation of secondary very iron-rich biotites is encountered in some of the metasomatized granitoids (Chatterjee and Strong, 1984).

WHITE MICA

The occurrence of primary white mica in peraluminous granites demands pressures generally in excess of three kb., depending on the composition of the mica (Miller et al., 1981; Anderson and Rowley, 1981). The parageneses of white mica in the SMB are legion, including secondary after biotite, cordierite, andalusite, K-feldspar and plagioclase, but nowhere has it been possible to demonstrate unequivocally a magmatic origin for white mica (Ham, 1983). In some cases it is possible to statistically discriminate between the white micas of the various parageneses, but in no case do those suspected of having a primary magmatic origin on textural grounds define a unique compositional grouping.

ANDALUSITE

Andalusite occurs in intermediate- to late-stage rocks of the SMB. Its general absence from the marginal granodiorites, coupled with its apparent textural and chemical equilibrium with the other granitoid minerals, provide convincing evidence of a magmatic origin (Clarke et al., 1976). The conclusion regarding a magmatic origin demands use of the andalusite-sillimanite inversion curve of Richardson et al. (1969) and provides an upper pressure limit on the depth of crystallization of the SMB at about four kb, assuming there is no magmatic muscovite.

GARNET

Garnet is a very minor mineral in most phases of the SMB, but elsewhere in granitoids of the Meguma Zone, it can occasionally be very abundant (10-15% mode). Allan and Clarke (1981) described two principal types of garnets from the eastern part of the SMB: first - anhedral, low Mn, inclusion-rich varieties in reaction relationship with biotite and interpreted as relict xenocrysts from the country rocks; second - euhedral higher Mn, inclusion-free varieties with no reaction relationship with biotite which are interpreted as being magmatic in origin. The xenocrystic types are found only in the marginal granodiorites of the SMB where they were clearly out of equilibrium with the melt phase, but the later stage magmatic varieties indicate that the melt did become saturated in garnet during crystallization of the batholith, as a result of either a higher A/CNK ratio or $Mn/(Mn+Fe)$ ratio in the magma or both.

CORDIERITE

Although not as modally abundant in the SMB as in the neighbouring Musquodoboit Batholith (MacDonald, 1981), the marginal granodiorites and monzogranites of the SMB still contain small amounts (<1%) of cordierite. Its most common mode of occurrence is as large (up to two cm), isolated, euhedral to subhedral crystals in the granitoid matrix. Maillet and Clarke (1985) have shown there to be a continuum in texture and chemistry from the cordierites of the thermal aureole rocks, through those of the xenoliths to those occurring as isolated grains in the granitic rocks, and have interpreted this to indicate that at least some of the cordierites in the granitoid rocks have a metamorphic precursor. However, since the cordierites are euhedral (unlike aureole occurrences) and appear to have been in chemical equilibrium with the melt, additional direct crystallization from the melt cannot be precluded. Furthermore, cordierites occur in some late-stage pegmatites of the SMB where presumably it belongs to the 'hyperaluminous' stage of crystallization (Clarke, 1981) in which the presence of a fluid phase has been instrumental in creating melts or fluids with high A/CNK ratios.

OTHER CHARACTERISTIC MINERALS

Other minerals which reflect the peraluminous or hyperaluminous nature of the SMB rocks include tourmaline, which is a common phase in both pegmatites and greisens, topaz which is the product of hydrothermal alteration in the presence of fluids containing high concentrations of fluorine (Muecke and Clarke, 1981; Logothetis, 1984), and a garnet-rich spinel associated with greisen zones (Farley, 1979). These minerals largely owe their existence to the interaction of fluid phases and either magma or solid rocks. Since the altered rocks have $A/CNK > 1$ it is deduced that the complementary fluid phases were peralkaline ($A/CNK < 1$) in nature and were preferentially removing alkalis. The presence of albitized rocks in such places as East Kemptville reflects the deposition of some of these alkalis.

Oxide Phases

Ishihara (1977) has recognized that the occurrence of ilmenite is a distinctive feature of the peraluminous granites of Japan. This association is also true for the SMB, although primary oxide minerals are extremely scarce in all phases of the batholith. Logothetis (1981) has described ilmenites with compositions $(\text{Ilm}_{97}\text{Pyroph}_3 - \text{Ilm}_{77}\text{Pyroph}_{23})$ which are generally less Mn-rich than ilmenites of the country rocks. In addition, Dwyer (1975), Charest (1976) and Logothetis (1981) have all described occurrences of members of the columbite-tantalite series as well as niobian rutile as trace primary accessory phases in intermediate- to late-stage phases of the SMB and other plutons of the Meguma Zone.

GEOCHEMISTRY

The broad petrologic spectrum of the SMB granitoids is reflected in their geochemistry,

with SiO_2 values generally ranging from 68-76%. Averages for the major and selected trace elements of the non-metasomatized rock types are presented in Table 1-1. Many of the major elements and trace elements show smooth trends when plotted on variation diagrams (for examples see Fig. 1-2). Statistically these are reflected by strong positive and negative correlation coefficients (Logothetis, 1984). In such a grouping, element correlations are noticeably poor for the leucomonzogranites and highly fractionated rocks. Chatterjee et al. (1983) pointed out that in these late-stage rocks of the SMB two geochemically distinct suites (Sn-specialized and U-specialized) can be identified on the basis of their trace element chemistry. The U-specialized suite is particularly marked by elevated Rb, Zn, Li and P_2O_5 values, whereas the Sn-specialized suite is preferentially enriched in Sn, U and Th. Chatterjee and Muecke (1982) distinguished these two suites on the basis of their U and Th content, and demonstrated that their distribution within the SMB can be delineated by airborne radiometric surveys.

The peraluminous character of the SMB is confirmed by the prevalence of corundum-normative compositions and a general increase in normative corundum with differentiation index (McKenzie and Clarke, 1975). Hyperaluminous compositions are confined to rock types which have undergone significant fluid interaction. A notable anomaly in element behaviour is evident in the lanthanides. In mafic systems they behave as incompatible elements and show a systematic increase with fractionation. In the SMB and other peraluminous granitoids (Miller and Mittlefehldt, 1982; Mittlefehldt and Miller, 1983; Price, 1983), their abundance decreases with differentiation. Muecke and Clarke (1981) demonstrated that this is particularly marked in the late-stage rocks. Further studies of mineralized and metasomatized lithologies in the SMB have documented even more pronounced depletions (Logothetis, 1984; Chatterjee and Strong, 1984).

The continuous and smooth element variation trends in the SMB strongly suggest a cogenetic origin of all the rock types. McKenzie and Clarke (1975) concluded that crystallization of plagioclase and biotite from a parental granodiorite magma could adequately account for most of the chemical variation. Subsequent isotopic and geochemical work essentially concurs with their conclusions but indicates that other processes must also have played a significant role. This is particularly true for the highly differentiated rocks where elemental trends show the influence of aqueous fluids interacting with the magma or solidified rocks. For example, Muecke and Clarke (1981) invoked the action of fluorine-rich solutions to explain the observed lanthanide depletions. Studies on SMB greisens and metasomatized rocks support this hypothesis.

Stable isotope compositions of oxygen and sulphur provide another constraint on the possible source regions of the magmas of the SMB. Longstaffe et al. (1980) determined $\delta^{18}\text{O}$ values of +10 to +12 ‰ for the SMB, but found no significant differences between the different

Table 1-1. Average major and selected trace element composition of the various granitoid rocks from the SMB. (Analytical data from McKenzie and Clarke, 1975; Muecke and Clarke, 1981; Farley, 1979; Charest, 1976; and Chatterjee et al., 1983). The number of samples in the average (n) varies from element to element for a given rock type.

	Granodiorites n = 4-11			Monzogranites n = 5-30			Leucomonzogranites n = 3-25			Dike Bodies (Aplites, Pegmatites, Elvans) n = 4-12			Greisens (Turner Tin, Walker Moly) n = 2-10			Tin-specialized Granite n = 94			Uranium-specialized Granite n = 136		
	$\bar{x}$	s.d.		$\bar{x}$	s.d.		$\bar{x}$	s.d.		$\bar{x}$	s.d.		$\bar{x}$	s.d.		$\bar{x}$	s.d.		$\bar{x}$	s.d.	
SiO ₂	69.38	1.60		74.06	1.21		74.45	1.16		73.78	2.26		72.95	16.10		75.83	2.43		74.48	0.97	
TiO ₂	0.61	0.12		0.20	0.09		0.07	0.06		0.07	0.05		0.06	0.03		0.13	0.12		0.08	0.01	
Al ₂ O ₃	14.36	0.32		13.62	0.39		14.06	0.51		14.73	1.37		15.43	9.34		12.65	1.37		13.50	0.55	
Fe ₂ O ₃	0.44	0.18		0.16	0.06		0.21	0.11		0.24	0.17		0.56	0.39		-	-		-	-	
FeO	3.53	0.72		1.51	0.35		0.81	0.22		0.72	0.19		1.75	1.63		1.21	0.31		0.93	0.08	
MnO	0.10	0.01		0.04	0.02		0.03	0.01		0.03	0.01		0.93	0.04		0.04	0.05		0.05	0.06	
MgO	1.02	0.21		0.28	0.16		0.10	0.09		0.10	0.08		0.14	0.05		0.18	0.11		0.05	0.06	
CaO	1.93	0.28		0.59	0.19		0.51	0.14		0.46	0.11		0.46	0.18		0.67	0.27		0.75	0.14	
Na ₂ O	3.29	0.18		3.45	0.27		4.11	0.60		3.98	0.67		0.89	1.30		3.41	0.83		3.48	0.27	
K ₂ O	3.86	0.27		4.72	0.31		4.32	0.49		4.95	1.70		5.26	2.82		4.08	0.82		4.39	0.38	
P ₂ O ₅	0.10	0.04		0.14	0.07		0.26	0.11		0.24	0.13		0.22	0.03		0.11	0.03		0.44	0.04	
H ₂ O	0.88	0.13		0.80	0.16		0.84	0.12		0.81	0.10		1.97	1.26		0.76	0.23		0.63	0.13	
F	0.07	0.01		0.15	0.07		0.40	0.32		0.27	0.20		0.70	0.43		0.33	0.10		0.45	0.06	
Ba	689	75	181	124	124	57	104	57	89	57	89	44	25	73	60	39	24				
Sr	140	24	25	15	15	14	11	15	10	14	10	18	14	31	112	38	13				
Rb	142	16	388	132	132	653	317	760	264	653	264	879	329	700	122	769	63				
Li	124	53	277	140	140	559	583	628	334	559	334	-	-	146	64	517	48				
Sn	12	3.7	22	9.74	44	40	18	44	20	40	20	-	-	50	60	39	8				
W	2	-	1	0.00	2	-	2	2	-	-	-	-	-	11	4	2	5				
Mo	2.5	0.58	2.0	0.82	1.7	0.7	0.7	1.7	0.5	1.5	0.5	-	-	2	0.1	1.4	0.6				
Cu	11	4.80	8	5.12	13	5	13	13	1	5	1	-	-	7	26	31	25				
Zn	68	8	60	9.42	55	39	23	55	6	39	6	-	-	43	14	86	22				
Th	10.94	1.90	10.61	5.15	3.94	2.68	1.40	3.94	2.05	2.68	2.05	4.41	0.24	25	6	5	2				
Cs	11.01	2.93	23.61	6.09	34.07	23.73	10.99	34.07	6.21	23.73	6.21	61.71	23.10	-	-	-	-				
Sc	10.67	2.26	3.68	0.98	2.49	2.06	1.19	2.49	0.47	2.06	0.47	2.55	0.95	-	-	-	-				
Ta	1.81	0.54	3.55	1.55	6.84	6.59	2.35	6.84	3.86	6.59	3.86	4.90	1.84	-	-	-	-				
Hf	5.58	1.34	3.01	0.77	1.92	1.56	0.23	1.92	0.74	1.56	0.74	2.08	1.62	-	-	-	-				

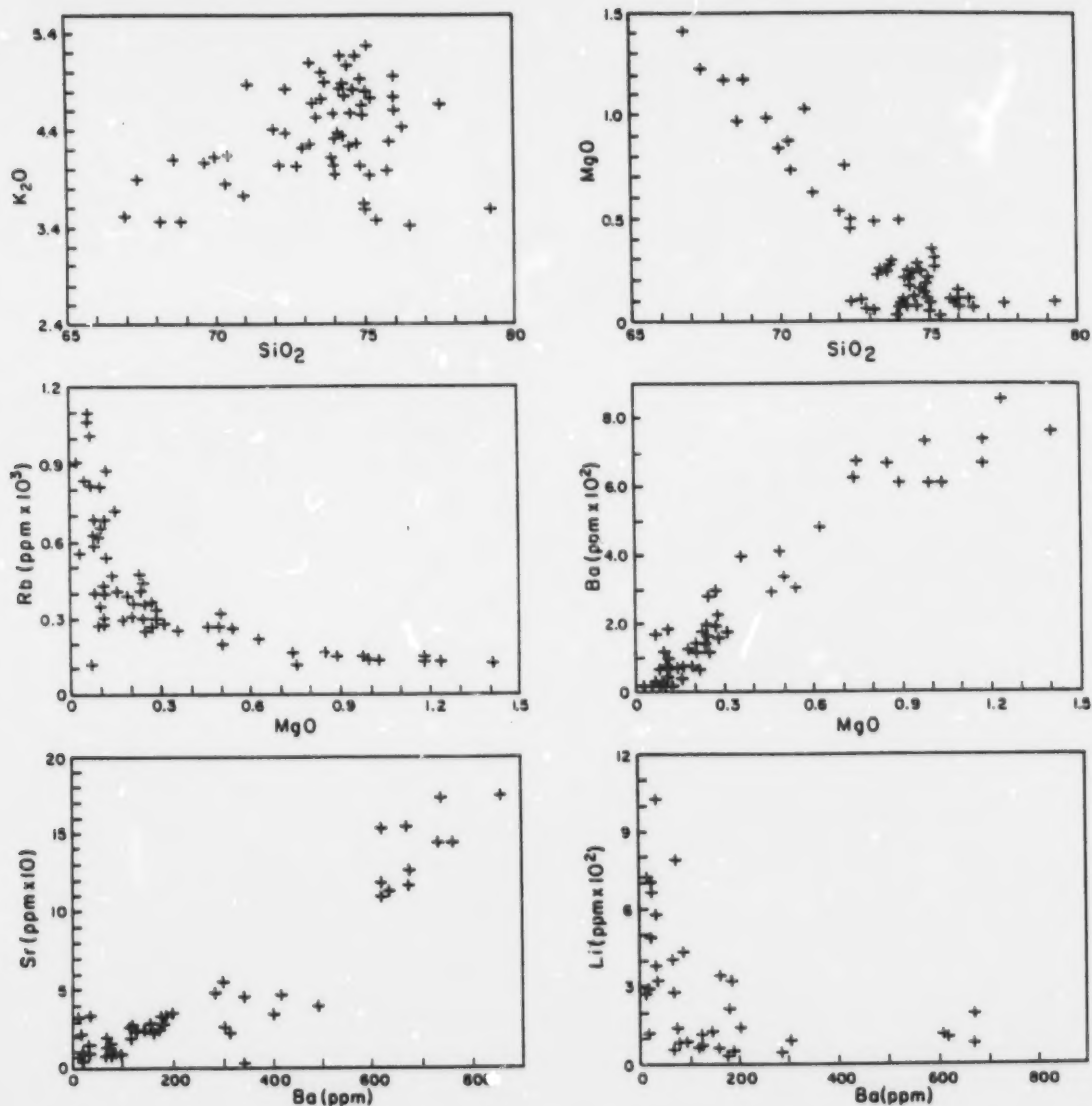


Figure 1-2. Geochemical variation in the South Mountain Batholith. Various types of correlation are: SiO_2 - K_2O , linear positive correlation between major elements; SiO_2 - MgO , linear negative correlation between major elements; MgO - Rb , exponential negative correlation between major and trace elements; MgO - Ba , linear positive correlation between major and trace elements; Ba - Sr , linear positive correlation between trace elements; and Ba - Li , exponential negative correlation between trace elements.

phases. Meguma metasediments yielded a range of $\delta^{18}\text{O}$ from +9 to +13 ‰ and their partial melting could yield granitoids with $\delta^{18}\text{O}$ equivalent to those of the SMB. Recent work on the mineralized and metasomatized granitoids shows both positive and negative shifts in the oxygen isotopic ratios, possibly reflecting the interaction with meteoric as well as internally generated fluids (Muecke et al., 1985). Sulphur isotopic ratios range from $\delta^{34}\text{S}$ +5.4 to +8.0 ‰ in the granodiorites and show a wider range from +1.6 ‰ to +15.0 ‰ in the monzogranites (Kubilewicz and Ohmoto, 1984). Since the Meguma metasediments range from +9.6 ‰ to +18.6 ‰ at least the granodiorites cannot have derived their sulphur from a Meguma source.

RADIOGENIC ISOTOPIC STUDIES

The granitoid rocks of the Meguma Zone have been the object of a considerable amount of geochronological and isotope geochemical work. Reynolds et al. (1984) have summarized all the relevant K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ work and have shown that the mean age of 23 samples from the SMB is 366.7 ± 4 Ma, but were not able to detect any statistical difference in the mean ages of the granodiorites and later intrusives. Metasomatized and altered granitoids associated with tin and uranium mineralization have yielded distinctly younger argon ages of about 280 Ma (Zentilli and Reynolds, 1985; Reynolds et al., 1981).

Early Sr isotopic studies by Fairbairn et al. (1964) and by Cormier and Smith (1973) are of limited value due to inadequate knowledge of the intrusive relations in the SMB and large analytical errors. Clarke and Halliday (1980) showed that three main intrusive phases could be defined and that the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio increased with each episode, as follows: granodiorite -372 ± 2 Ma, 0.7081; monzogranite -364 ± 1 Ma, 0.7094; and late-stage aplites and leucomonzogranites -361 ± 1 Ma, 0.7102. In addition, they showed that the contact facies granodiorites were isotopically contaminated with country rock, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the extensive Meguma envelope was too high to be the sole source of the peraluminous granitoids, and the increase in $^{87}\text{Sr}/^{86}\text{Sr}$ could be accounted for by a closed system fractionation model, but that additional contamination from the country rocks between episodes was likely.

Neodymium isotopic studies on the metasediments of the Meguma Group show that they have a mean crustal residence age of 1.77 ± 0.95 Ga, and that they had ϵ_{Nd} values of -9 to -13 at the time of intrusion of the granites (Clarke and Halliday, 1985). Further work on the granites by Clarke et al. (1985) shows that the ϵ_{Nd} decreases from the early granodiorite to the latest leucomonzogranite, the xenoliths have ϵ_{Nd} values intermediate between country rock and granodiorite, and the ϵ_{Nd} values for the granodiorite are higher than those of the Meguma flyschoid rocks. The implication is that either a mantle component is present in the batholith, or the source region has a younger mean crustal residence time than the Meguma Group.

MINERALIZATION

The SMB is the host of a recently discovered major tin deposit at East Kemptville (Richardson et al., 1982; Chatterjee and Strong, 1984b) and a uranium deposit at Millet Brook (Chatterjee et al., 1982; Chatterjee and Strong, 1984a) as well as numerous showings of Sn-W-Cu-Mo and U-Li-F-P associations. Mineralization is accompanied by intensive alteration and metasomatism of the granitoids. The deposits and occurrences include mineralized shear zones (eg. Millet Brook), aplite-pegmatite associations (eg. Walker, Morley's), various types of greisen (eg. East Kemptville, Long Lake), and hydrothermal veins (eg. Canoe Lake). Almost invariably these occurrences are directly associated with the most highly differentiated members of the SMB, or are suspected to be underlain at shallow depth by such a body. The majority of occurrences are concentrated in the vicinity of the two major radiometric and Bouguer anomalies which mark the site of the highly differentiated New Ross-Vaughan Complex and Davis Lake Pluton.

Many of the occurrences have been studied in detail and the reader is referred to the following additional sources for further information: Charest (1976), Farley (1979), Logothetis (1984), O'Reilly and Chatterjee (1981), O'Reilly et al., (1982).

SUMMARY AND CONCLUSIONS

The SMB is a posttectonic, peraluminous biotite granodiorite-two-mica monzogranite - leucomonzogranite complex which includes highly 'specialized' granitoids and is the host to significant mineralization. The major intrusive event of the batholith can be bracketed by isotope geochronology and stratigraphy to within a 10-12 Ma interval. Unroofing of at least part of the batholith by Tournaisian time implies rapid uplift during or after this event. Locally the hydrothermal history of the complex appears to have been much more prolonged. Primary mineral parageneses and stratigraphy confine pressures at the time of intrusion to one to four kb (three to ten km depth) at $T < 850^\circ\text{C}$. Continuous and smooth element variation trends suggest that all members of the complex are part of a cogenetic suite.

Regardless of the model one adopts, the least differentiated biotite granodiorites provide the closest approximation to the parental material and are already peraluminous in composition. However, it is not clear whether this excess alumina is of anatectic, contamination or alteration origin. Isotopic, trace element and phase relations evidence all point to a largely metasedimentary source, but cannot exclude the possibility of a limited mantle component. An anatectic origin from Meguma Group metasediments is feasible from the oxygen isotope data. However, the strontium, neodymium, and sulphur isotope ratios, as well as the lanthanide abundances, all preclude the Meguma Group as the sole source. Some attributes of the source region can be defined. It would apparently be characterized by ^{18}O -rich rocks (e.g. mature siliclastic sediments), lower Rb/Sr or younger

age than the Meguma Group, lower lanthanide abundances and $\delta^{15}\text{S}$ ratios than the Meguma Group, and neodymium ratios characteristic of material with a younger mean crustal residence time. At the present time it is not possible to resolve whether these chemical characteristics are those of lower crustal metasedimentary rocks alone, or whether a mantle component is also required.

ITINERARY

START: 0.00 KM/0.00 MI LORD NELSON HOTEL

Proceed north along South Park Street and Bell Road and west along Quinpool Road to MacDonald Street just before the bridge over the railway tracks.

STOP 1: 2.74 KM/1.70 MI* METASEDIMENTARY ROCKS OF THE MEGUMA GROUP (FIG. 1-3)

(NOTE: This locality can only be visited with prior permission from Canadian National Railways).

The Appalachians have been divided into several lithotectonic zones (Williams, 1978), all but one of which can be traced for considerable distances along the orogen stretching from Newfoundland to Alabama. Only the Meguma Zone, the outermost of the zones, is anomalous in that it cannot be correlated longitudinally anywhere else in the orogen. It is characterized by two main lithologies, namely a thick sequence of flyschoid rocks metamorphosed and deformed during the Acadian orogeny, and a number of post-tectonic peraluminous granitoids.

The flyschoid rocks have been divided into two main lithologies, namely a metawacke and a metapelite. Where the overall sand/shale ratio is greater than unity, the material has been mapped as the Foldenville Formation and, where the ratio is less than unity, as the so-called Halifax Formation. Schenk (1982) has interpreted these as continental slope and rise proximal and distal turbidite deposits, respectively. Indeed, most single units show at least part of the typical Bouma sequence from the massive sandstone

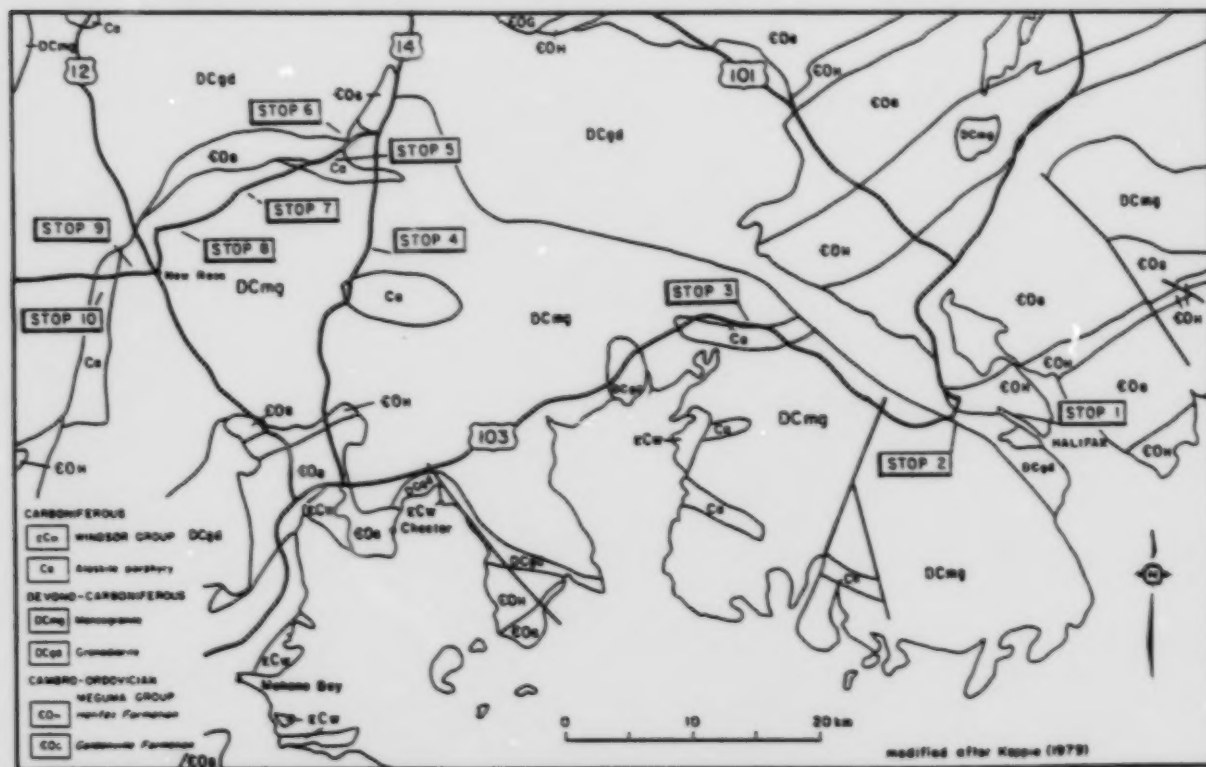


Figure 1-3. Map showing the geology and field trip stops in the New Ross-Halifax area, southwestern Nova Scotia.

*Cumulative distances outbound from Halifax on main routes only and do not include short side trips off the main roads.

(A) division, to the parallel laminated (B) division, to ripple laminated (C) division, to the silty (D) division, to the uppermost pelitic (E) division. Primary sedimentary structures such as load casts, flutes, grooves, channels, and intraformational clasts, are abundant. However the Meguma Group is essentially unfossiliferous.

Our interest in these metasedimentary rocks is that they form nearly the entire envelope of the peraluminous granites of the Meguma Zone at its present level of erosion. To a first approximation, their chemical composition is rather restricted, and this in turn puts limits on the possible effects of contamination of the granitic magmas as they intruded. In fact, the chemical composition of the metawacke is very similar to the granodiorites (see Clarke and Chatterjee, 1985; and Clarke, Muecke and Chatterjee, this volume). In the vicinity of Halifax these rocks have been raised to greenschist facies of regional metamorphism and, at this locality, we are within the contact aureole of the Batholith and therefore andalusite or cordierite (or both) of thermal origin may also be observed. We will see these rocks again as enclaves in the granodiorite at the next stop, but at this locality we should focus on the lithological variation within single units (graded bedding) and on the style of deformation, which is as steep, isoclinal folds everywhere in the Meguma Zone, undisturbed by the intrusive event.

Continue west along Quirpool Road to the Arndale Rotary. Take the St. Margaret's Bay Road exit from the Rotary and follow the signs to Highway 103. Stop at the underpass of Highway 103 just before the entrance ramp into the highway.

STOP 2: 7.4 km/4.6 MI MARGINAL GRANODIORITE OF THE SMB

Exposed here is the typical pristine slightly porphyritic biotite granodiorite (qx-kapar-plag-biot-ap-zircon $\pm$ musc: there is no amphibole in the batholith) of the contact of the SMB. It is the earliest and most abundant rock type in the batholithic complex. Close to known contacts, this phase contains abundant enclaves of country rock, usually identifiable as metawacke similar to material seen at Stop 1.

This exposure is cut by several saccharoidal aplites and, although hydrothermal alteration is slight in the granodiorites, there is still evidence of a separate vapour phase in the pegmatite dykes found here. The pegmatites are interesting in that they contain large crystals of both tourmaline and cordierite, further evidence of the highly peraluminous nature of the Batholith.

Take the left turn onto the ramp onto Highway 103 towards Yarmouth and proceed to a prominent outcrop on the main highway at the top of the hill 0.8 km/0.5 mi past the Upper Tentation/Hammond Plains Road/Peggy's Cove exit (Exit 5).

STOP 3: 31.4 km/19.5 MI CORDIERITE-RICH MONZOG-RANITE

For a full description of this rock type, refer to MacDonald (this volume). Briefly, this phase of the SMB is unusual because of its large modal abundance of cordierite, the origin of which is almost certainly different from the pegmatitic cordierite at Stop 2. In this case, it is probably either purely magmatic in origin, or has been derived in part from the metapelites of the contact aureole.

Also at this locality are many fractures filled with fluorite. Based on the high REE concentrations in some of these fluorites, Muecke and Clarke (1981) have suggested that fluorine may have been important as a complexing agent in the mobilization of the rare earth elements in the late stages of crystallization of the SMB. Occurrence of fluorite is common in the SMB, and it will be seen at several other stops on our trip.

Continue 35.0 km/21.7 mi west along Highway 103 to Exit 8. Turn north on Highway 14 towards Windsor. Travel up this road about 13.5 km/8.4 mi to an outcrop on the right side near the top of a hill (2.9 km/1.8 mi beyond Armstrong Road).

STOP 4: 79.5 km/49.4 MI TWO-MICA MONZOG-RANITE AND ANDALUSITE APLITES

At this locality we are well within the New Ross-Vaughan Complex (NRVC), the largest of the predominantly monzogranitic bodies which are intrusive into the marginal granodiorites of the SMB (see Clarke, Muecke and Chatterjee, this volume). The monzogranite host exposed here has more biotite, and is therefore darker in colour, than most members of the NRVC. In addition, muscovite also occurs but its origin (primary or secondary) is not clear. There is also some modal cordierite in the monzogranite. A number of brown-weathering aplite dykes also occur here and contain from 2-10% modal andalusite, however whether the andalusite occurs in aplites or in the coarser grained monzogranites (elsewhere), it is never discernable in hand specimen. (In other words, if you don't believe this assertion, take a sample and get a thin section made of it). Clarke et al. (1976) have used spatial, textural and phase equilibrium arguments to conclude that the andalusite is a primary mineral in the NRVC, and then used the primary andalusite as a rough geobarometer to give a maximum pressure of about 3-4 kb for emplacement of the SMB.

Continue along Highway 14 towards Vaughan and Windsor. At 25.6 km/15.9 mi from Highway 103 notice the strongly reddened, hydrothermally altered monzogranites on the right side of the road. We are not going to take the time to stop here, but only to mention that pink and red altered granitoids are much more common in the interior of the batholith than at its margin. Although we did see evidence of fluids at the earlier stops (pegmatites and fluorite veins), their activity was largely confined to fractures, whereas near the centre of the batholith the effects are much more pervasive. At 29.6 km/18.4 mi from Highway 103, turn left at Vaughan on the

road towards New Ross. Travel along this road 3.5 km/2.2 mi and stop at a low outcrop on the left side of the road.

STOP 5: 99.4 KM/61.8 MI CANOE LAKE LEUCOMONZOG-GRANITE

At this locality may be seen one member of what is believed to be the latest, and chemically most evolved magmatic rocks of the NRVC (see Clarke, Muecke and Chatterjee, and Charest et al., this volume, for details). This rock type is unmistakable in the Batholith because of its low modal abundance, or even absence, of mafic minerals, moderate amounts of muscovite and close association with fluorite. This is not our best exposure of leucomonzogranite because of its somewhat mafic and, in parts, porphyritic character. Time permitting, we will be able to see better examples at Lewis Lake (STOP 7), and especially at the Reeves tin deposit (STOP 10). We only stop at this particular spot for two reasons: (i) to observe a new lithology, and (ii) because it has been suggested by Charest (1976), Charest et al. (this volume), and Logothetis (this volume) that major mineral deposits of the NRVC are spatially and perhaps genetically related to bodies of leucomonzogranite. This particular body is closely related to the Millet Brook uranium deposit which we will examine at the next stop (STOP 6).

Turn the vehicle(s) around and backtrack 0.5 km/0.3 mi to a road on the left that leads to the Millet Brook uranium deposit. Travel up this road a distance of 4.0 km/2.5 mi.

STOP 6: 99.9 KM/62.1 MI MILLET BROOK URANIUM DEPOSIT

The U - Cu $\pm$ P $\pm$ CO₂ $\pm$ B mineralization at Millet Brook is distributed in closely spaced pods and lenses confined to northeast trending shear and fracture zones in the 370 Ma old biotite granodiorite, the dominant and the least differentiated rock type of the Batholith. The mineralization has been shown by Chatterjee and Strong (1984b) to be genetically related to later, more differentiated uraniumiferous leucongranites which have a high percentage of leachable uranium.

The unaltered granodiorite is characterized by quartz, zoned plagioclase (An₃₃), perthitic K-feldspar (Or₅₀Ab₁₀), biotite (F/M = 2.04), muscovite, and accessory Mn-rich garnet, apatite, zircon, pyrite, pyrrhotite, ilmenite and rutile.

The unaltered granitoid rocks are chemically homogeneous and similar to that of the Batholith in general, excepting a few elements which show a greater variance such as Ba, Cu, Rb, Zn and Li. There is a general increase in variance from the unaltered through the altered to the mineralized granodiorite, especially for SiO₂, CaO, CO₂, Cu, Sr, Zn and U. By means of detailed statistical evaluation of the chemical data for 128 samples, using discriminant analysis and R-mode factor analysis, Chatterjee and Strong (1984b) were able to show that different processes were responsible for the chemical variation in each of these groups.

Alteration within the granitoid rocks is generally confined to within 50 meters of the mineralized fractures and includes potassic, sodic, calcic and ferruginous varieties, with resulting secondary K-feldspar, albite, biotite, muscovite, chlorite and calcite. The rare minerals are characterized by pitchblende, torbernite, autunite and Pb-meta-autunite, together with minor amounts of chalcopryrite, bornite, covellite, chalcocite, proustite, galena, sphalerite, saite and wolframite. The associated gangue minerals are quartz, feldspars, micas, illite-smectite, andalusite, hematite, kaolinite, tourmaline, chlorite, calcite-ankerite, pyrite, Mn-oxide and anatase.

The altered granitoid rocks can be subdivided into two main groups, one dominated by the effects of aqueous hydrothermal fluids and the second by carbonate-rich fluids. The latter 'carbothermal' or 'episyenite' alteration produces a distinct geochemical signature marked by concentration of elements Sn, Cu, U, Pb, Co, Ca and Mn and to a lesser extent F, Be and Th. It is suggested that these elements were transported as carbonate complexes in the hydrothermal solutions and that their deposition resulted from exsolution of CO₂ from this fluid and its reaction with the host rocks, in a manner similar to that suggested by Coney (1978) and Leroy (1978) for the similar uranium deposits of the French Central Massif.

Along with the carbonate alteration, there are other significant types, such as silicification, hematization, albitization and K-feldspathization/sericitization, presumably reflecting the complexity of these hydrothermal fluids. The chemical effects of these alteration processes can be seen in Figure 1-4 which shows the different types trending away from the unaltered granodiorite. This diagram reveals something of the nature of the processes, in that the trends are not directed toward the particular mineral involved, but more away from or towards the most mobile elements of silica and potassium. For example, the albitized rocks trend toward albite + quartz, reflecting more a loss of the K-feldspar component than simple addition of albite, the episyenites trend more away from silica, and the silicified rocks trend directly toward silica, reflecting only the mobility of silica. Figure 1-4 also shows the distribution of quartz and feldspars in the mineralized samples, with no apparent restriction to rock type.

At the Millet Brook deposit, the alteration and associated uranium mineralization will be examined at surface (trenches on C₁ and C₂ zones) and at depth in drill core (DDH A9-1 and A9-23) from the Upper Salter Lake Zone. The location of the drillholes in plan and cross-section is illustrated in Figure 1-5. The details on the deposit and the mineralogical and geochemical characteristics of the host rocks are described by Chatterjee et al. (this volume, Chapter 5).

Turn the vehicle(s) around again and travel west in the direction of New Ross for a distance of 8.9 km/5.5 mi to a rather low, flat, inconspicuous outcrop on the left side of the road near the top of a hill.

STOP 7: 108.8 km/67.6 MI LEWIS LAKE LEUCOMONZOG-GRANITE

At this locality there occurs a qz-ksp-peg-musc-fluorite leucomonzogranite free of mafic minerals. (Is all the fluorite of hydrothermal origin, or is some of it magmatic?) In some cases it is suspected that the leucomonzogranites are the source of the mineralizing fluids (examples of this close relationship are the Millet Brook uranium deposit just visited, the Walker Sn-W-Mo-Cu deposit to be visited next, and the Reeves Sn deposit probably not visited on this trip but where the deposit is hosted entirely within a pristine leucomonzogranite body).

Continue west along the New Ross Road for a distance of 5.0 km/3.1 mi stopping just beyond a small hill and house on the right at a road on the left side of the road. The Walker deposit is a short stroll down the road toward the house and then turn right at the house and walk approximately 100 m along the road to the deposit.

STOP 8 113.8 km/70.7 MI WALKER POLYMETALLIC SN-W-MO-CU DEPOSIT

The only outcrop consists of a small exposure of leucomonzogranite near the old shaft, so the nature of the deposit itself has to be inferred from old reports and from material in the dump. The main lithologies are a banded aplite-pegmatite consisting of qz-felds-musc-molybdenite-copper sulphides, and a coarse grained gneiss consisting of qz-musc-siderophyllite-topaz-tour-m-fluorite-molybdenite-copper sulphides-tennantite ($\text{Cu}_{12}\text{As}_2\text{Sb}_2\text{S}_{13}$) - mawsonite (CuFeSnS_6) - stannite ($\text{Cu}_3\text{FeSnS}_4$) - wittichenite (Cu_3SbS_4) - uraninite-rutile-gahnite-wolframite-native bismuth. Deposition seems to have extended over a temperature interval from at least 610° C to 270° C, i.e., from the magmatic to the low temperature hydrothermal range. Further details on this deposit may be found in Charest et al. (this volume) and Farley (1979).

Proceed 6.0 km/3.8 mi along the road to the junction of Highway 12 in New Ross. Depending on time available, the trip may conclude here, or two optional localities may be visited. Stop 9 introduces one further important lithology (porphyry), and Stop 10 provides a review of several of the features examined so far. From New Ross proceed west along the Forties Road for a distance of 4.0 km/2.5 mi to the D.M. Reeves property.

STOP 9: 123.9 km/77.0 MI LAKE RAMSAY MONZO-GRANITE PORPHYRY

Monzogranite porphyries are just textural variants of the other monzogranites found in the NRVC (Charest et al., this volume). The pink porphyry at this locality is characterized by phenocrysts of Ksp + Qz set in a fine grained groundmass. This rock type is very common in this part of the complex and, despite the fact that the porphyries are nearly barren of sulphides, many of the mineral deposits of this area have incorrectly been referred to as 'porphyry coppers'.

Continue driving west along the Forties Road a further 1.0 km/0.6 mi to some houses at the top of a hill. Turn left and walk or drive to the farmhouse at the end of the road.

STOP 10: 124.9 km/77.6 MI REEVES' TIN DEPOSIT

The road to the farmhouse begins in biotite granodiorite, but about half way along the road a major contact between the outer envelope of granodiorite and two-mica monzogranites of the NRVC is crossed. By the time we have reached the deposit (which is located 100-200 metres southwest of the farmhouse), we have crossed another contact (not exposed) with the Reeves-Keddy leucomonzogranite. This beautiful leucomonzogranite is much better exposed than at the Walker deposit, contains small disseminated blue crystals of fluorite, and is host for a mineralized pegmatitic segregation or dyke. The open pit itself, the pegmatite and much of the dump have recently been covered up in the name of safety.

Minerals reported from this deposit are topiolite ($\text{Fe(Ta,Nb)}_2\text{O}_6$), cassiterite, lepidolite, amblygonite, apatite, fluorite, durangite ($\text{NaAl}_2(\text{As},\text{F})_2$), columbite, morinite ($\text{Ca}_2\text{Na}_2\text{Al}_4(\text{PO}_4)_2\text{F}_6\cdot 5\text{H}_2\text{O}$), monazite, topaz, beryl, tourmaline, metatorbernite, wolframite, scheelite and tungstate. Most of the mineralized samples are no longer collectable.

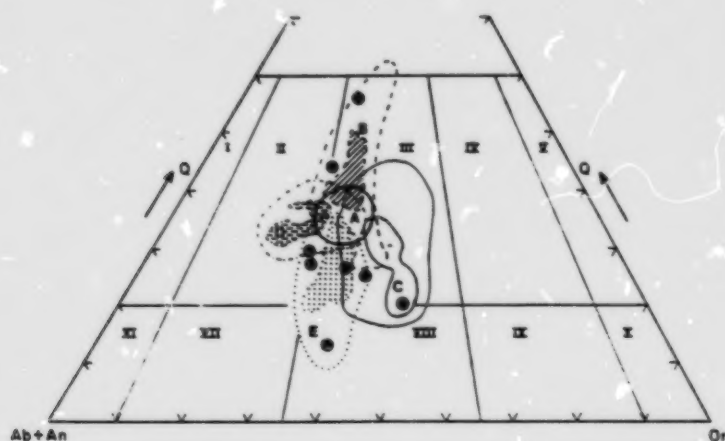


Figure 1-4. Normative Q-Ab+An-Or plot showing the distribution of unaltered rocks (field A), silicified rocks (field B), K-feldspathized rocks (field C), albited rocks (field D) and desilicified rocks: 'episyenites' (field E) from the intra-batholith uranium deposit at Millet Brook (N=179). The patterned areas within the designated fields represent approximately 75% of the data points. Arterisks jacketed by a circle represent the compositional range of the mineralized rocks.

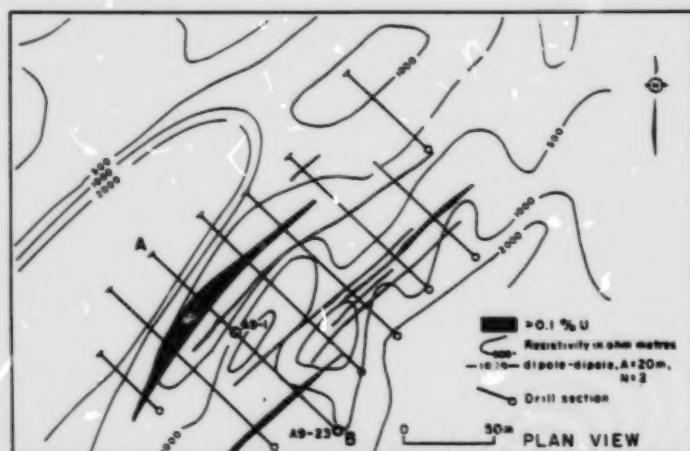
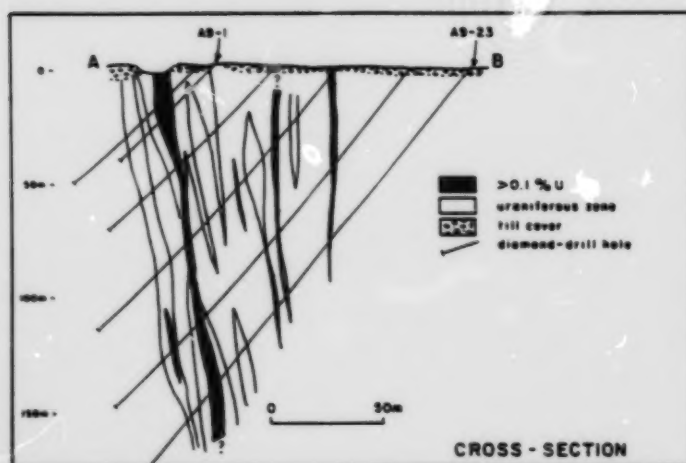


Figure 1-5. Plan and cross-section showing the location of the drillholes at the Upper Salter Lake zone: Millet Brook area.



CHAPTER 2

A TRAVERSE THROUGH THE SOUTHEASTERN PART OF THE SOUTH MOUNTAIN BATHOLITH: THE GRANITE GEOLOGY OF HIGHWAY 103 BETWEEN HALIFAX AND CHESTER

M. A. MACDONALD

MacDonald, M. A., A traverse through the southwestern part of the South Mountain Batholith: the granite geology of Highway 103 between Halifax and Chester; in Guide to the Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

INTRODUCTION

In recent years numerous authors have successfully subdivided large areas of granitic rocks into separate mappable units by using both textural and compositional criteria (Cobbing et al., 1981; Dickson, 1983). The purpose of this contribution is to record some of the typical variations in a portion of the South Mountain Batholith (SMB) prior to the commencement of a major mapping program in the Batholith and to determine if these techniques for classification can be applied in the SMB as a whole. The excellent exposure of granitoid rocks along Highway 103 was chosen as the study area. The eastern end of the Highway from St. Margaret's Bay to Halifax crosses the Halifax Peninsula previously mapped by Smith (1974). Smith recognized a late intrusive 'alaskite' body near Tantallon and, with the exception of granodiorite near the Halifax granite/metasediment contact, mapped the rest of the rocks as quartz monzonite which he later referred to as biotite granite (Smith and Turek, 1976).

RESULTS OF FIELD MAPPING

In the fall of 1984, one week was spent mapping the field area. Thirty-six representative samples were collected (Fig. 2-1 for locations), slabbed, and stained for alkali feldspar content using sodium cobaltinitrite. A thin section was prepared from each sample. The granitoid rocks were subdivided into five lithotypes on the basis of their textures and mineralogy (Fig. 2-1). Classification of rock types is after Streckeisen (1976). Large euhedral, perthitic alkali feldspar phenocrysts commonly in excess three cm long are termed megacrysts. Grain sizes are defined as: fine < 0.1 cm; medium 0.1 to 0.5 cm; and coarse 0.5 to 3 cm. All rocks contain quartz, alkali feldspar and plagioclase. Biotite, when present, contains inclusions of apatite and zircon unless otherwise stated. Saussuritization refers to the secondary minerals produced by deuteric or hydrothermal alteration of calcic plagioclase. These include zoisite, calcite, sericite and calcite. Pinite is the common alteration product of cordierite and "consists of a fine felty mixture of muscovite with some chlorite or serpentine minerals and iron oxides" (Deer et al., 1966).

Unit 1 Biotite Granodiorite (Fig. 2-2A): medium grained; light bluish grey to grey; equigranular with perthitic megacrysts; megacrysts (< 5% of outcrop) frequently are so inclusion-rich that they may be confused with groundmass; biotite (approximately 15%) is fresh; plagioclase displays oscillatory zoning with slightly to moderately saussuritized calcic cores and may have apatite inclusions; trace muscovite occurs as intergrowths with biotite (secondary?) and as alteration of alkali feldspar; trace opaques (ilmenite?); metasedimentary xenoliths and aplite dykes are common.

Unit 2a Biotite Transitional Monzogranite: Medium- to predominantly coarse-grained; light grey to grey; generally equigranular with minor K-feldspar megacrysts (< 5%); compositions on QAP (quartz-alkali feldspar-plagioclase) diagram cluster near the granodiorite-monzogranite join; biotite is predominantly fresh with minor

chlorite alteration and may have inclusions of zircon and apatite in zones; muscovite is present in trace amounts, but is slightly more abundant than in the biotite granodiorite; muscovite is intergrown with biotite, but is predominantly an alteration of alkali feldspar; plagioclase is normally zoned with some development of oscillatory zoning, calcic cores may be moderately saussuritized; anhedral to subhedral, interstitial cordierite is a minor phase (< 1%) and is totally pseudomorphed by pinite and lesser amounts of muscovite; plagioclase may contain discrete inclusions of apatite and biotite; metasedimentary xenoliths are common.

Unit 2b Biotite Transitional Monzogranite (Fig. 2-2B): predominantly medium grained matrix with larger, coarse grained quartz, alkali feldspar and plagioclase; brownish- to greenish-grey; inequigranular, hypidiomorphic granular and massive (nonfoliated); slightly megacrystic (< 5%); compositions straddle monzogranite-granodiorite join on QAP diagram; biotite is relatively fresh with minor alteration to chlorite plus acicular rutile, apatite and zircon inclusions may be zoned; biotite may also contain stubby, subhedral rutile inclusions; trace muscovite as in Unit 2a; subhedral interstitial cordierite is partially to totally pseudomorphed by pinite plus minor muscovite; plagioclase is normally zoned (oscillatory zoning in part) with slight to moderate saussuritization of calcic cores; metasedimentary xenoliths are common.

Unit 3a Biotite Monzogranite (Fig. 2-2C): medium- to predominantly coarse-grained, bluish grey to grey hypidiomorphic granular; megacrystic (up to approximately 20% of outcrop); biotite is fresh with minor chloritization; inclusions of apatite are larger than in Units 1-2b (< 1 mm versus < 0.1 mm); trace to minor muscovite occurs mostly as alkali feldspar alteration and to a lesser extent in the cores of zoned plagioclase; cordierite may have fresh cores with sector twinning (Fig. 2-2D), however it is extensively altered to pinite, muscovite and chlorite; plagioclase has normal and patchy zoning with slightly saussuritized cores; hand specimens and unstained slabs resemble biotite granodiorite, however staining reveals substantial alkali feldspar content; thin aplite and pegmatite dykes (mostly < 10 cm) are abundant which may contain garnet, tourmaline or beryl.

Unit 3b Biotite-Muscovite Monzogranite (Fig. 2-3A): medium- to coarse-grained, medium greenish grey to grey, hypidiomorphic granular, megacrystic (5-20%); may have weakly defined foliation by parallel alignment of megacrysts; aplites present but not abundant; may have large (> 1 cm) subhedral to euhedral cordierite crystals (partially to totally pseudomorphed by muscovite, chlorite and pinite) which seldom constitute more than 2% of rock; fresh cordierite may be twinned; cordierite also occurs as subhedral interstitial totally pseudomorphed grains; biotite (5-8%) is fresh to extensively chloritized and may have stubby subhedral rutile crystals; plagioclase has normal zoning with extensively saussuritized cores; muscovite (> 1%) commonly replaces alkali feldspar in irregular patterns (Fig. 2-3B); quartz, alkali feldspar and plagioclase may poikilitically enclose small,

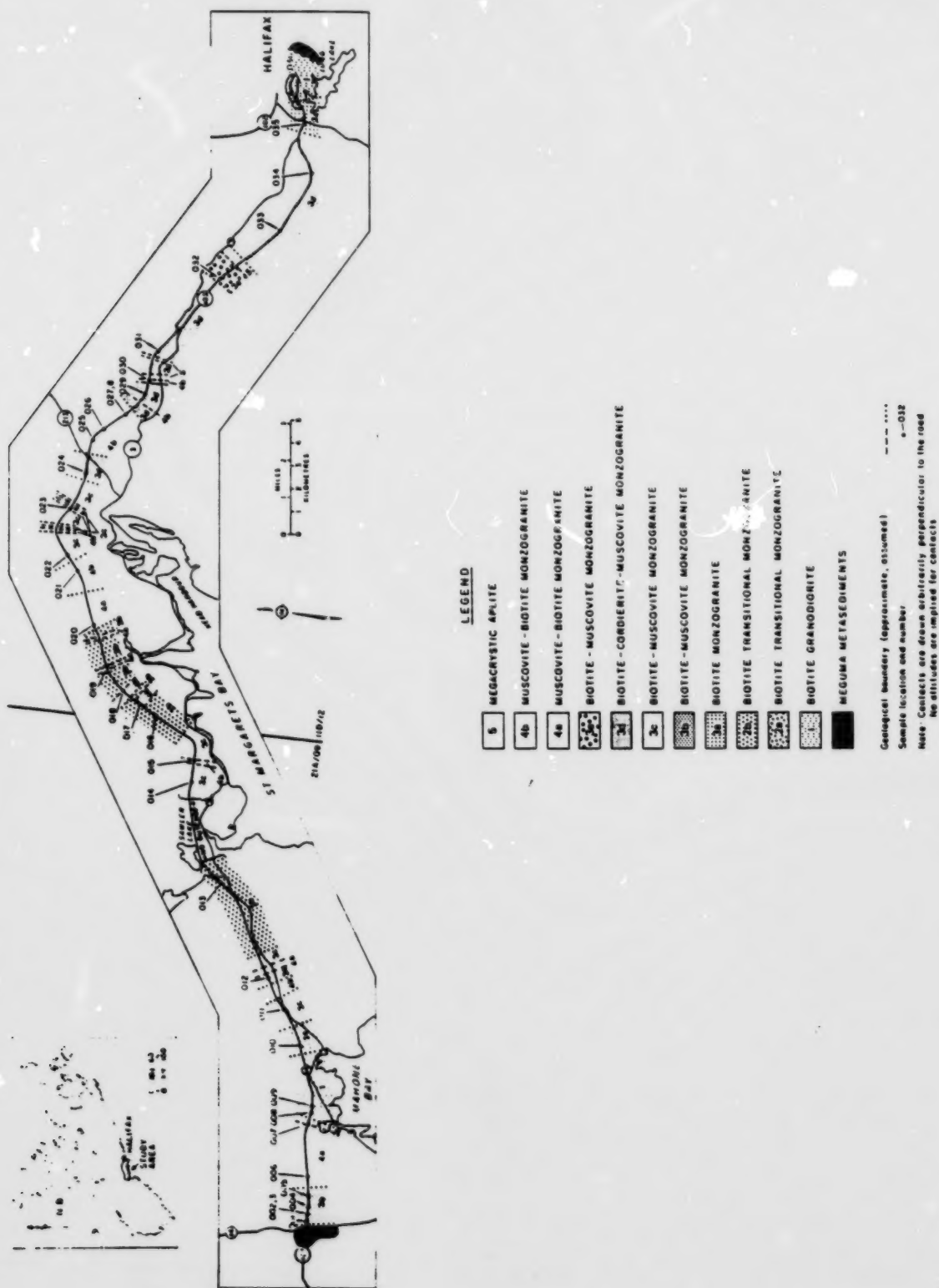


Figure 1. The granite geology of Highway 103 between Halifax and Chester.

fresh, inclusion free crystals of biotite; minor metasedimentary xenoliths.

Unit 3c Biotite Muscovite Monzogranite (Fig. 2-3C): fine- to medium-grained, coarse-grained in part; greenish- to brownish-grey; inequigranular, hypidiomorphic granular; megacrystic in part (< 5% of outcrop); massive biotite is generally fresh but frequently shows extensive chloritization and may have stubby subhedral rutile inclusions and ilmenite (along cleavage planes); alkali feldspar may be moderately sericitized; plagioclase shows normal zoning with variably saussuritized calcic cores; cordierite is present in minor amounts (1-2%) and occurs predominantly as small subhedral interstitial crystals with minor euhedral, blocky hexagonal prismatic crystals (> 1 cm long); cordierite is generally totally replaced by muscovite, chlorite and minor amounts of pinite; muscovite (1-5%) is mostly a replacement of alkali feldspar; may be devoid of xenoliths and aplite-pegmatite dykes in part.

Unit 3d Biotite Cordierite Muscovite Monzogranite (Fig. 2-3D): medium- to predominantly coarse-grained; light to medium grey, brownish- or bluish-grey; hypidiomorphic granular, inequigranular; very megacrystic (> 50% megacrysts in part, average 20-30%); large euhedral pseudo-hexagonal prismatic crystals of cordierite (0.5-1.5 cm long) are diagnostic (may constitute > 5% of mode); biotite may contain elliptical inclusions (parallel to cleavage) of alkali feldspar (and ilmenite?) similar to features described in the Musquodoboit Batholith (MacDonald, 1981); cordierite is invariably altered to pinite, muscovite and chlorite; a few euhedral muscovite crystals were noted, and these may represent primary magmatic crystallization; megacrysts may contain inclusions of quartz in zones; tourmaline and garnet (Fig. 2-4A) may be present as accessory minerals; outcrops are commonly foliated (Fig. 2-4A) with both pervasive parallel arrangements of megacrysts and highly irregular orientations; metasedimentary xenoliths and aplite-pegmatite dykes (< 10 cm thick) are common features.

Unit 3e Biotite Muscovite Monzogranite (Fig. 2-4B): medium grained, some outcrops have fine grained matrix (approaching aplitic); medium yellowish- to brownish-grey; equigranular, hypidiomorphic granular, megacrystic in part (5-15%); megacrysts are yellowish orange in colour and are inclusion-rich with inclusions mostly of quartz in zones; secondary muscovite is present as very small- to medium-sized irregular crystals and approaches equal proportions with biotite in part; clusters of complexly intergrown chlorite and muscovite may represent the pseudomorphic replacement of cordierite; biotite is extensively altered to chlorite; plagioclase is mostly unzoned, however, some is normally zoned with extensively altered cores, a few zircon inclusions were noted in plagioclase; massive, nonfoliated; minor metasedimentary xenoliths.

Unit 4a Muscovite Biotite Monzogranite: fine- to medium-grained; yellowish to orangish-brown; equigranular with minor amounts of quartz, alkali feldspar, plagioclase and muscovite phenocrysts, ranging in size from 0.5 to 1 cm and constituting < 5% of rock; muscovite occurs as secondary

alteration products as in other units and as primary-looking, subhedral to euhedral crystals; alteration of biotite ranges from fresh to extensively chloritized $\pm$ opaque mineral (ilmenite?); biotite may have stubby rutile inclusions, but is characteristically inclusion-free; plagioclase is typically unzoned or slightly zoned (albitic composition) and is slightly altered to saussurite $\pm$ calcite; euhedral to subhedral primary andalusite similar to that reported by Clarke et al. (1976) (i.e. always rimmed by secondary muscovite) and apatite are accessory mineral phases; massive (i.e. non-foliated); fractures in outcrop usually contain manganese or hematite staining.

Unit 4b Muscovite Biotite Monzogranite: fine- to medium-grained, grain size tends to be highly variable; wide range of textural characteristics from equigranular with a few phenocrysts to porphyry textures; these textural and grain size variations may vary drastically over several tens of metres (e.g. Figs. 2-4C and 2-4D); phenocrysts range in size from 0.5 to 2.0 cm and are primarily comprised of quartz, alkali feldspar, plagioclase (often with albite rims) and smaller phenocrysts (< 0.5 cm) of muscovite and biotite; muscovite is mainly an alteration product of alkali feldspar, however, minor euhedral primary-looking crystals are present; biotite is usually chloritized to some degree and some grains contain stubby subhedral rutile inclusions; phenocrysts of plagioclase display normal or patchy zoning whereas smaller groundmass grains are mostly unzoned (albitic) or weakly zoned; euhedral, primary andalusite (< 1 mm), similar to that described in Unit 4a, may be present in trace amounts; myrmekite intergrowths of quartz and plagioclase, and quartz and alkali feldspar may occur; complex intergrowths of muscovite and chlorite similar to those in Unit 3e may represent pseudomorphs after cordierite.

Unit 5 Megacrystic Aplite (this unit is limited in areal extent (Fig. 2-1) and the description is based entirely upon field observations): fine grained matrix with common megacrysts (up to 20%); light brownish grey, groundmass may be equigranular (aplitic) or slightly porphyritic with phenocrysts (< 1 cm) of quartz, alkali feldspar and plagioclase; may also have small (< 0.5 cm) biotite $\pm$ muscovite phenocrysts; proportions of biotite and muscovite are variable, however, biotite is usually the dominant mica.

SUMMARY AND CONCLUSIONS

It is apparent from this preliminary investigation that there is substantial textural and compositional variation in the granitic rocks of the SMB exposed along Highway 103. Parameters which may be used in subdividing the Batholith into lithotypes include: composition (i.e. feldspar ratio); grain size; colour; texture; mica content and proportion (i.e. biotite: muscovite); percentage of megacrysts; development of primary fabrics (i.e. flow foliation); content of cordierite and trace mineralogy (e.g. presence and mode of occurrence of andalusite, garnet, cordierite, tourmaline, rutile, apatite, beryl). A summary of the distinguishing characteristics of each unit is presented in Table 2-1.

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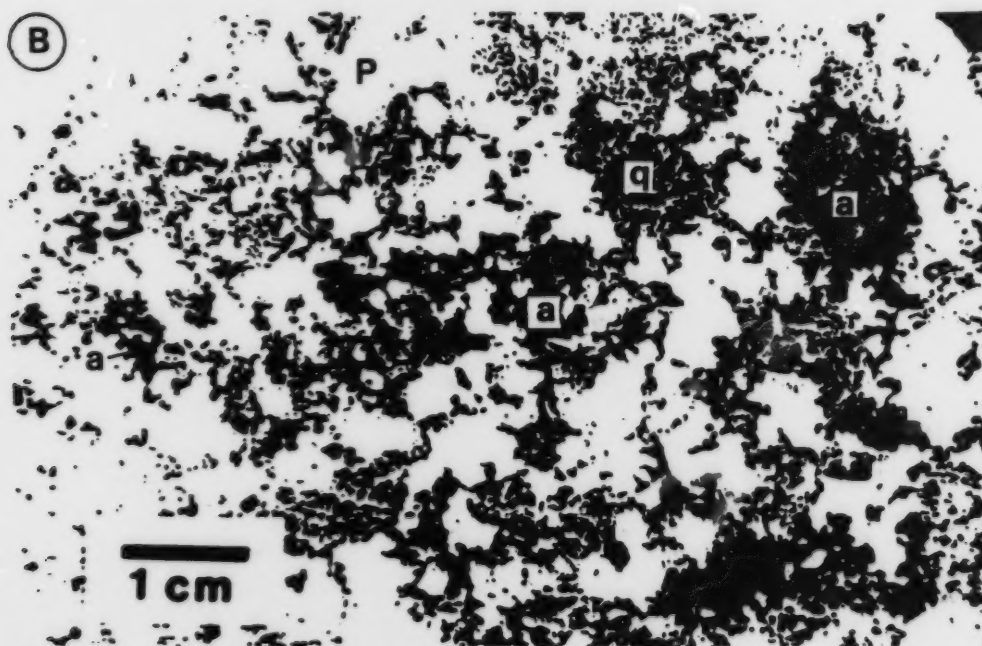
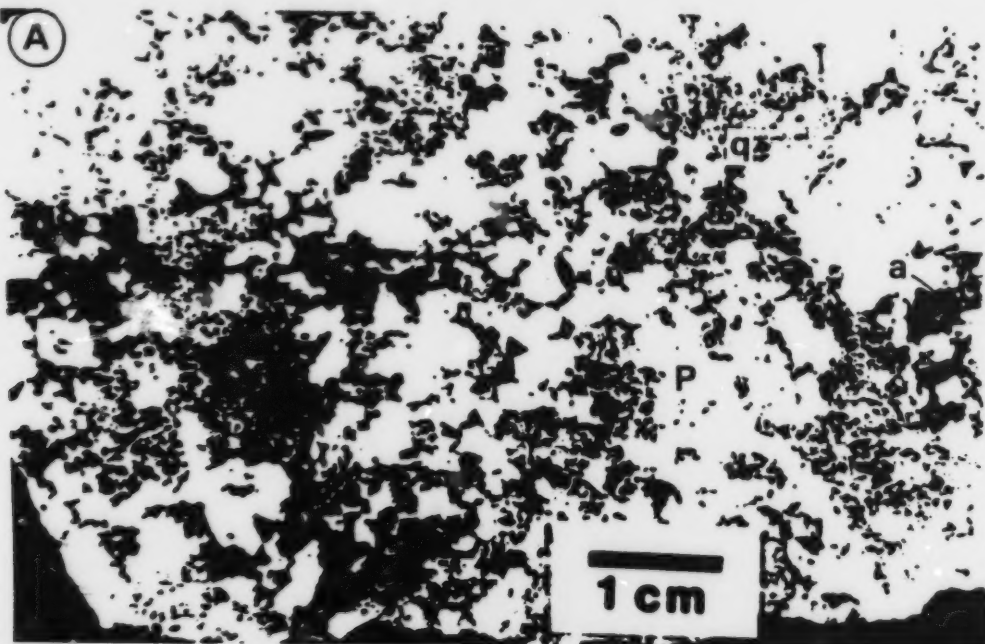


Figure 2-2. (A) biotite granodiorite (sample 036).

(B) biotite transitional monzogranite (sample 019).

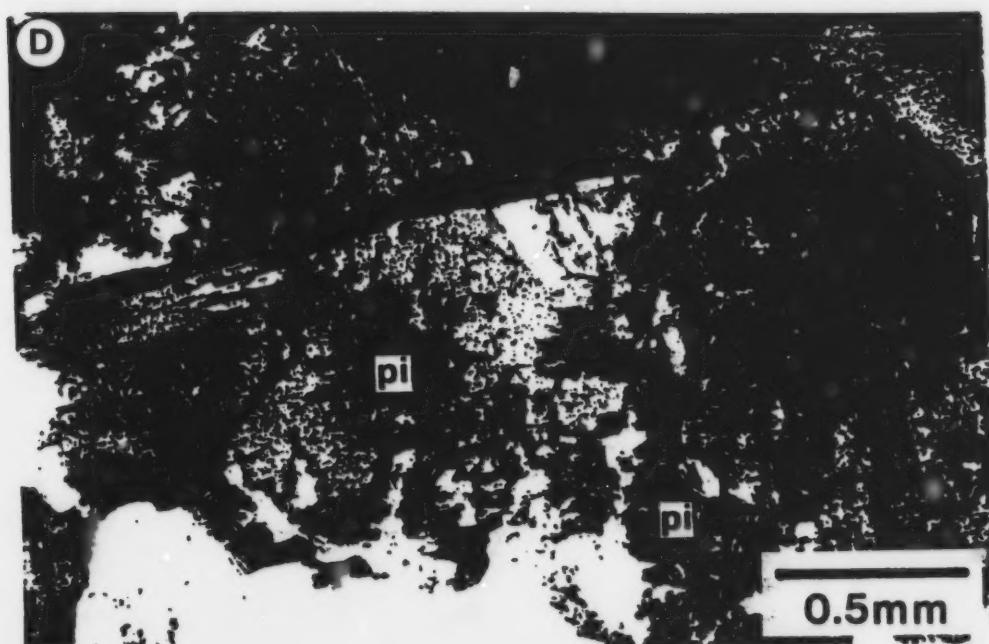
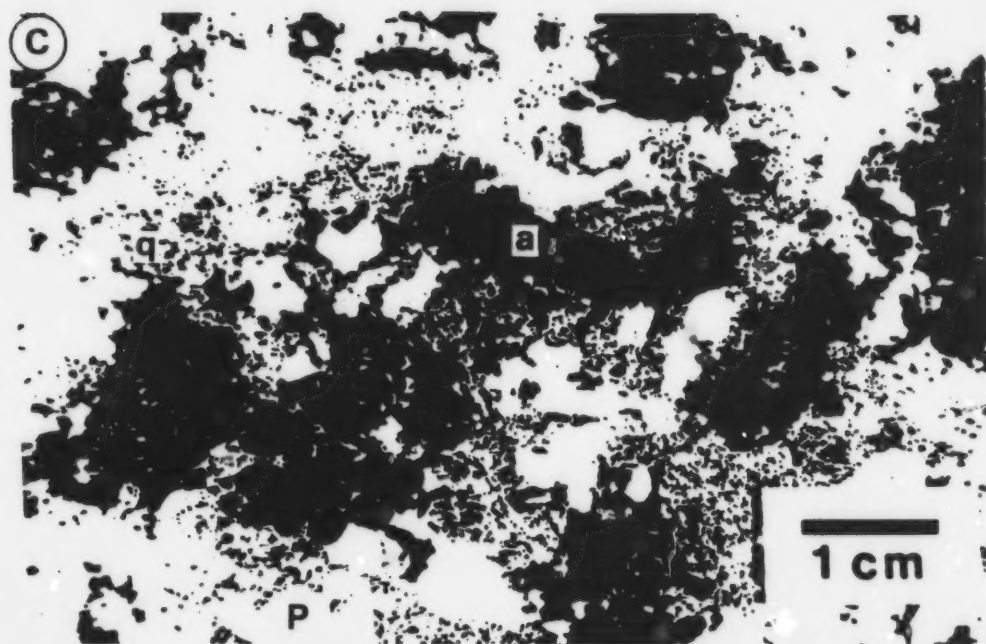


Figure 2-2. (C) biotite monzogranite (sample 035). Stained slabs show the proportions of light grey quartz (q), dark grey to black alkali feldspar (a) and white plagioclase (p).

(D) photomicrograph of cordierite (crossed polars) in biotite monzogranite (sample 035). Fresh cordierite in centre of photograph displays sector twinning. Note the characteristic alteration to pinite (pi).

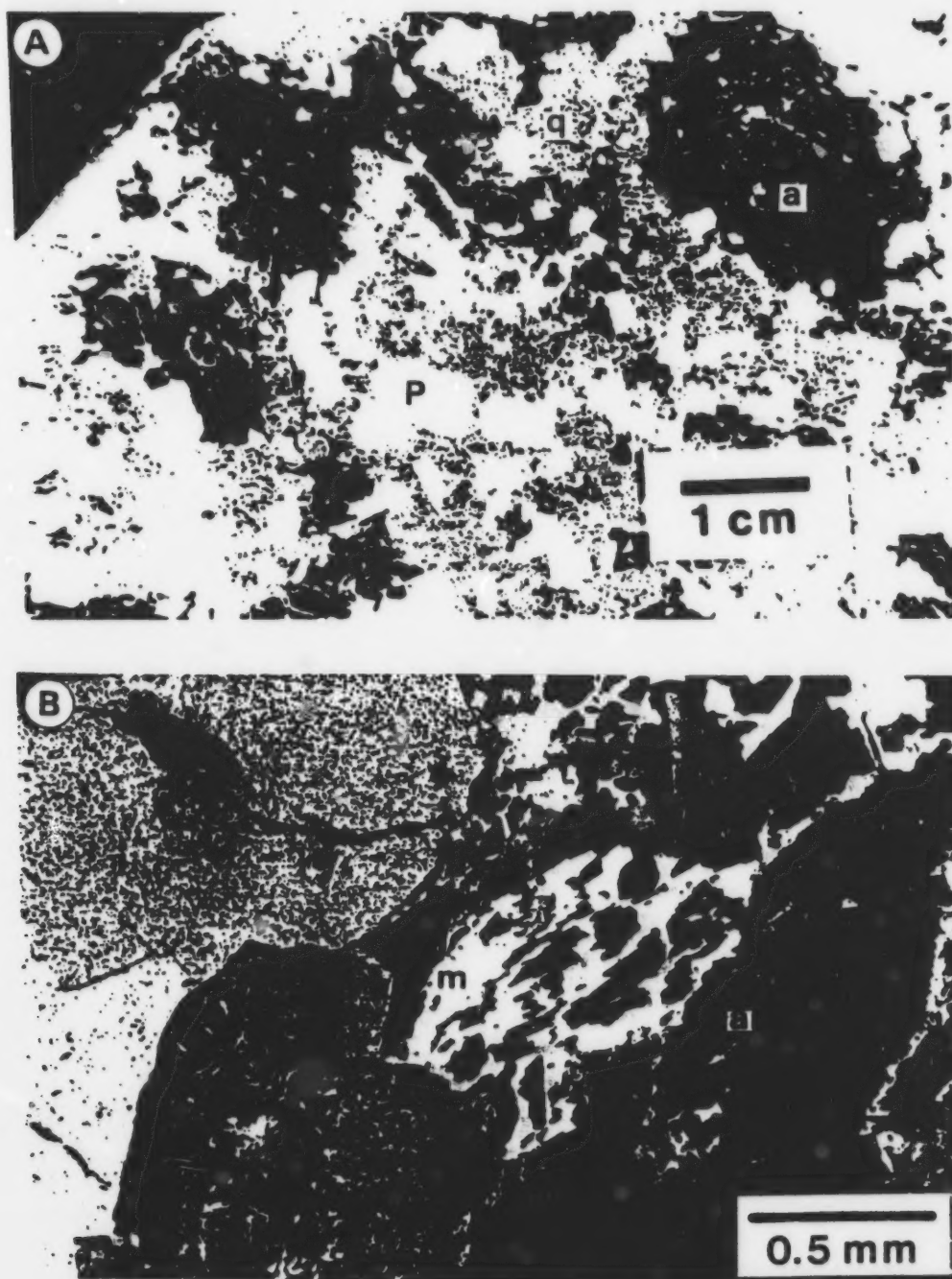


Figure 2-3. (A) Biotite-muscovite monzogranite (sample 005) stained slab. Large black alkali feldspars (a) display perthitic exsolution and plagioclase (p) and quartz (q) are whitish and light grey respectively.

(B) photomicrograph of muscovite (m) replacing alkali feldspar (a) (crossed polars) in biotite-muscovite monzogranite (sample 005).

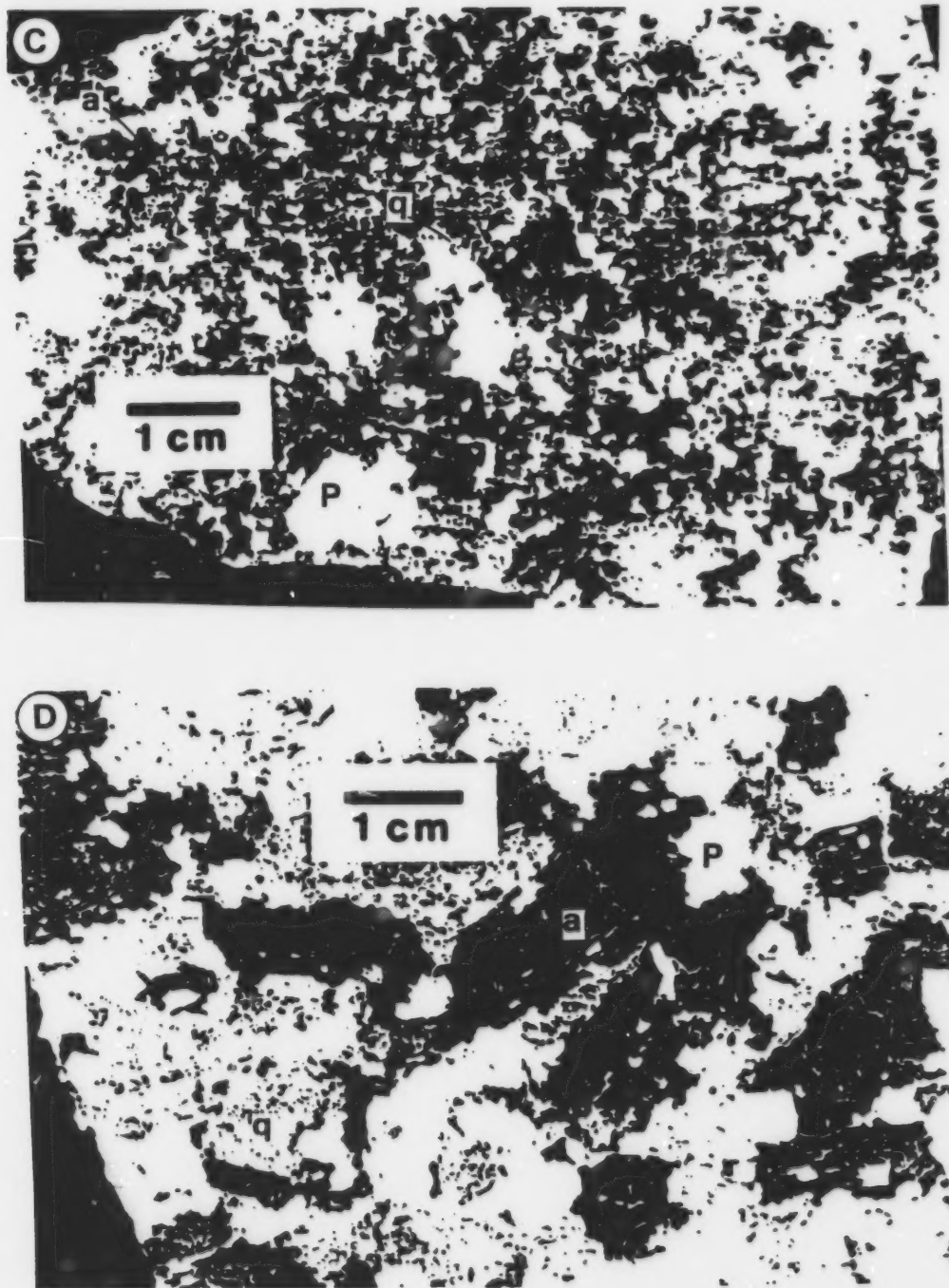


Figure 2-3. (C) biotite-muscovite monzogranite (sample 011).

(D) biotite-cordierite-muscovite monzogranite (sample 034) stained slabs show proportions of black alkali feldspar (a), whitish plagioclase (p) and grey quartz (q). Note the parallel arrangement of alkali feldspar megacrysts which defines a primary flow foliation.



Figure 2-4. (A) Photomicrograph (plane-polarized light) of a garnet inclusion (g) in plagioclase (p) in biotite-cordierite-muscovite monzogranite (sample 024). Note the anhedral crystal form which differs from the primary magmatic almandine-pegmatites. Note also the inclusion-free nature of the crystal which is unlike garnet from the metamorphic aureole.

(B) biotite-muscovite monzogranite (sample 032), unstained slab. Note the overall equigranular texture with a megacryst of alkali feldspar (a).

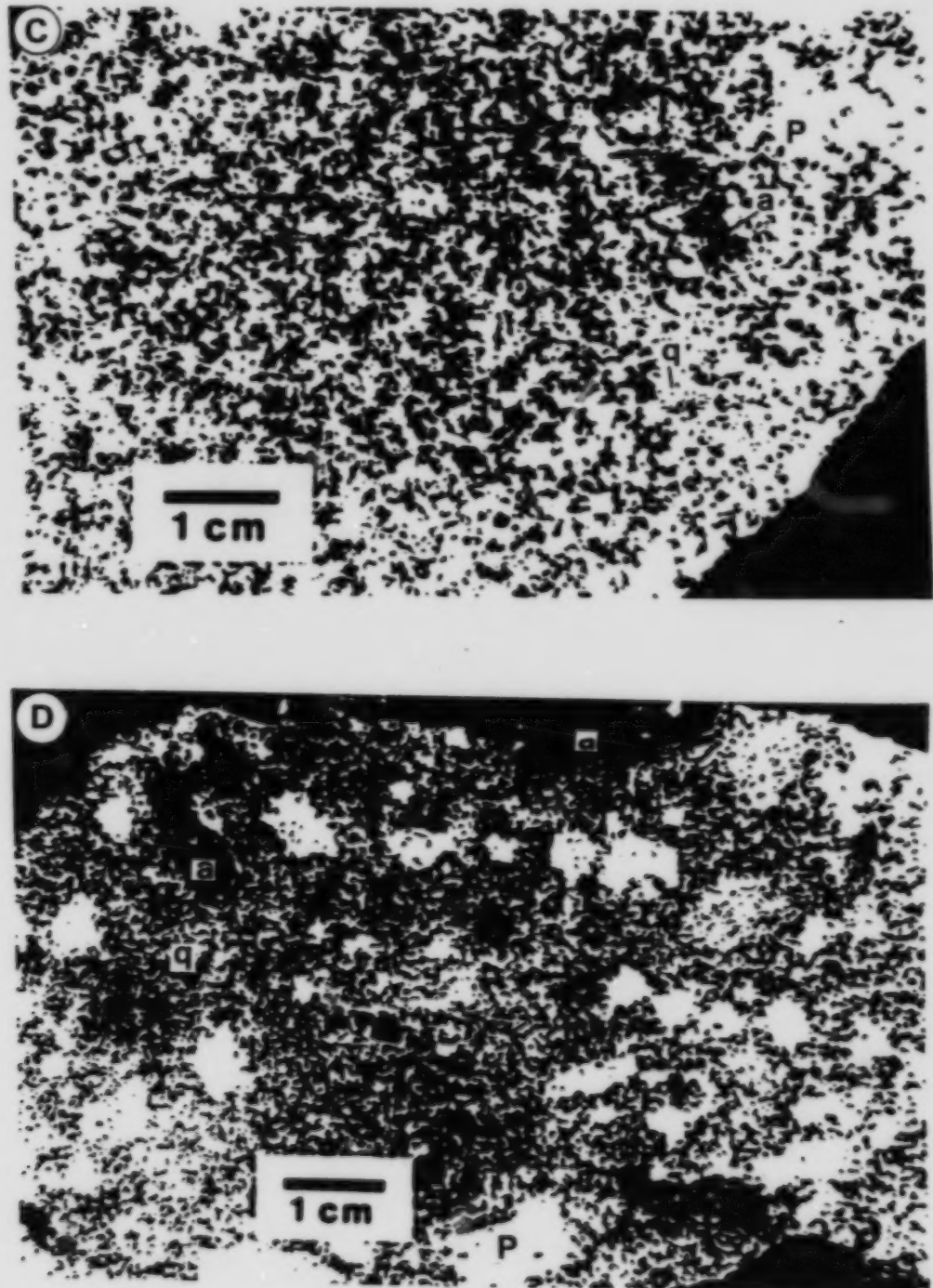


Figure 2-4. (C) and (D) Muscovite-biotite monzogranite (samples 027 and 028) stained slabs indicate proportions of black alkali feldspar (a), whitish plagioclase (p) and grey quartz (q). Samples were taken from the same unit (4b) approximately 30 m apart. Note the extreme variation in grain size and texture from fine grained equigranular (4C) to porphyritic (4D). Phenocrysts in sample 028 (4D) are comprised of quartz (q), alkali feldspar (a) and zoned plagioclase (p).

Table 2-1. Summary of distinguishing characteristics of each unit

UNIT NUMBER AND NAME	DISTINGUISHING CHARACTERISTICS
1 Biotite granodiorite	medium grained with $\leq 5\%$ megacrysts equigranular; light bluish-grey to grey; common metasedimentary xenolith; bi approximately 12-14%; common aplite dykes.
2a Biotite Transitional Monzogranite	monzogranite similar to Unit 1 except: medium- to predominantly coarse-grained; lacks bluish colour; more mu; $< 1\%$ ct (small subhedral crystals); more K-fel with compositions near granodiorite-monzogranite join.
2b Biotite Transitional Monzogranite	similar to Unit 2a except: predominantly medium grained; inequigranular.
3a Biotite Monzogranite	medium- to predominantly coarse-grained; bluish-grey to grey; common metasedimentary xenolith; trace mu and ct (may be fresh); resembles bi granodiorite, however stained slab indicates monzogranite composition; approximately 20% megacryst; common aplite dykes.
3b Biotite-Muscovite Monzogranite	medium- to coarse-grained; greenish-grey to grey; megacryst 5-20%; may be foliated; ct $< 1-2\%$ (as blocky crystals > 1 cm and smaller subhedral crystals); mu $> 1\%$; bi 5-8%.
3c Biotite-Muscovite Monzogranite	similar to 3b except: fine- to medium-grained; inequigranular; megacryst $\leq 5\%$; ct mostly small subhedral with minor large blocky crystals; may be devoid of xenoliths and aplite dykes.
3d Biotite-Cordierite-Muscovite Monzogranite	predominantly coarse grained; light- to medium-grey; megacryst up to 50%, mostly 20-30; foliated; large euhedral blocky ct up to 5%.
3e Biotite-Muscovite Monzogranite	medium grained; medium yellowish- to brownish-grey; locally matrix approaches aplitic; equigranular; mu $\leq$ bi.
4a Muscovite-Biotite Monzogranite	fine- to medium-grained; yellowish- to orangish-brown; equigranular with $\leq 5\%$ phenocrysts (0.5-1.0 cm); mu $\geq$ bi.
4b Muscovite-Biotite Monzogranite	similar to Unit 4a except: less equigranular; highly variable texture over tens of metres; very porphyritic in part $> 10\%$ phenocrysts (porphyry); phenocrysts (0.5-2.0 cm).
5 Megacrystic Aplite	similar to Unit 4b however majority of phenocrysts are K-fel megacryst in predominantly aplitic matrix.

List of abbreviations: bi - biotite; ct - cordierite; K-fel - alkali feldspar; mu-muscovite

CHAPTER 3

THE NORTHWESTERN PART OF THE NEW ROSS-VAUGHAN COMPLEX: PETROLOGY, GEOCHEMISTRY AND MINERAL DEPOSITS

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Charest, M. H., Farley, E. J. and Clarke, D. B., The northwestern part of the New Ross-Vaughan Complex: Petrology, geochemistry and mineral deposits; in Guide to Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

28 Granites & Mineral Deposits SW N.S.

ABSTRACT

The New Ross area is situated in the centre of the South Mountain Batholith (SMB) of Nova Scotia. It is part of a monzogranite pluton named the New Ross-Vaughan Complex (NRVC) which diapiroically intruded granodiorite. The rocks show a wide variation of texture and trace element chemistry, and a concentration of elements of economic interest (Sn, W, Mo, Cu, U). There is little major element chemical variation amongst the many styles and pulses of complex intrusion at New Ross. Some trace element (Sn, Ta, Ba, Sr, Zr, Rb) variations however, clearly demonstrate the magmatic differentiation sequence of monzogranite, monzogranite porphyry, aplite and leucocratic monzogranite. Concentration of other trace elements such as Li, Cu, Nb, Mo, W and Zn, are presumably due to processes involving aqueous fluids.

The emplacement of mineralized aplites, pegmatites, veins and greisens appears to be structurally controlled by the major contacts and by fractures, and petrologically associated with margins of very late-stage leucomonzogranites. It is believed however that orthomagmatic enrichment of such elements as Sn and Ta in leucomonzogranitic magmas has not led to any elemental concentration of economic interest.

Two mineral deposits have been studied in detail. The Walker deposit is a good example of transitional, or fluide-magmatic, enrichment. Molybdenite is associated with aplite and pegmatite, and other opaques such as wolframite, cassiterite, uraninite, chalcopryrite, bornite, gahnite, wittichenite, mawsonite and tennantite precipitated from greisenizing fluids. Emplacement of the deposit was in leucomonzogranite, at or near the contact.

The veins and greisens of the Turner deposit represent hydrothermal enrichment and were crystallized from deeper, water-saturated granitoid magma injected along fractures in the consolidated roof zone of the pluton. Complex opaque mineralization includes crystallization of arsenopyrite, pyrite, chalcopryrite, bornite, sphalerite, cassiterite, stannite and wittichenite.

INTRODUCTION

An overview of the South Mountain Batholith has been given by Clarke, Muecke and Chatterjee (this volume). Briefly, it consists of a discontinuous envelope of biotite granodiorite which has been intruded by its late differentiates as a number of discrete plutons of biotite-muscovite monzogranite. Rb-Sr isochron ages (Clarke and Halliday, 1980) show a progression from 372 Ma for the granodiorites, to 361 Ma for the youngest, and largest, of the differentiated plutons. This large pluton was named the New Ross-Vaughan complex (NRVC) by McKenzie (1974).

The western boundaries of the NRVC are well defined, but the positions of its other contacts are less well understood (Dwyer, 1976), although it is currently believed that rocks belonging to

this complex extend nearly to the eastern contact of the South Mountain Batholith. Even so, it is largely in the vicinity of New Ross that the complex shows a wide variation of texture and chemistry (but see MacDonald, this volume), and that there has been a concentration of elements of economic interest. (The tin deposit of East Kemptville, lying at the extreme southwest end of the South Mountain Batholith, and the uranium finds of Millet Brook and the Gaspereau Lake area, lying to the north of the NRVC, are discussed by Chatterjee et al., this volume). The purpose of this paper is to present mineralogical, petrological and chemical data for the host rocks and mineral deposits of the New Ross area, and to interpret the origin of these deposits in terms of the cooling and crystallization history of the South Mountain Batholith.

MAJOR LITHOLOGIES OF THE NEW ROSS-VAUGHAN COMPLEX

The area of interest lies at the northwestern extremity of the NRVC (Fig. 3-1), but is centered with respect to the principal regional gravity low (Garland, 1953; Douma, 1978). These studies indicate that there may be a thickness of up to 25 km of granitoid rocks in the area around the town of New Ross. A wide range of petrological, textural and chemical varieties of monzogranite compositions occurs within this area and represents many styles and pulses of intrusion. Exposures are poor, and from their variability, there appear to be many small plutons. These include small intrusions of coarse-grained monzogranite, monzogranite porphyry and leucocratic monzogranite, dykes of aplitic monzogranite and pegmatite, as well as veins and greisen zones, many of which are mineralized (see colour plates, to appear in final volume). These lithologies are summarized in Table 3-1 and further details may be found in McKenzie (1974), Charest (1976) and Farley (1979).

GEOCHEMISTRY

MAJOR ELEMENTS

Average chemical analyses for the principal rock types are given in Table 1-1 of Clarke, Muecke and Chatterjee (this volume). The granodiorites are clearly more basic than the other rocks of the Batholith, but in general there is little major element chemical variation amongst the monzogranites, porphyries, aplites and leucocratic monzogranites, presumably because the magma had reached an end-point in the system Qz-Kspar-Ab-An.

TRACE ELEMENTS

Table 1-1 (Clarke, Muecke and Chatterjee, this volume) also shows average trace element concentrations for the principal rock types of the South Mountain Batholith. The elements can readily be divided into two types, compatible, i.e. decreasing with differentiation (e.g. Ba, Sr, Zr), and incompatible, i.e. increasing with differentiation (e.g. Rb, Sn, Ta) with Cu, Mo, W and Nb not showing any strong trends except very close to some mineral deposits. The various types of positive and negative, linear and

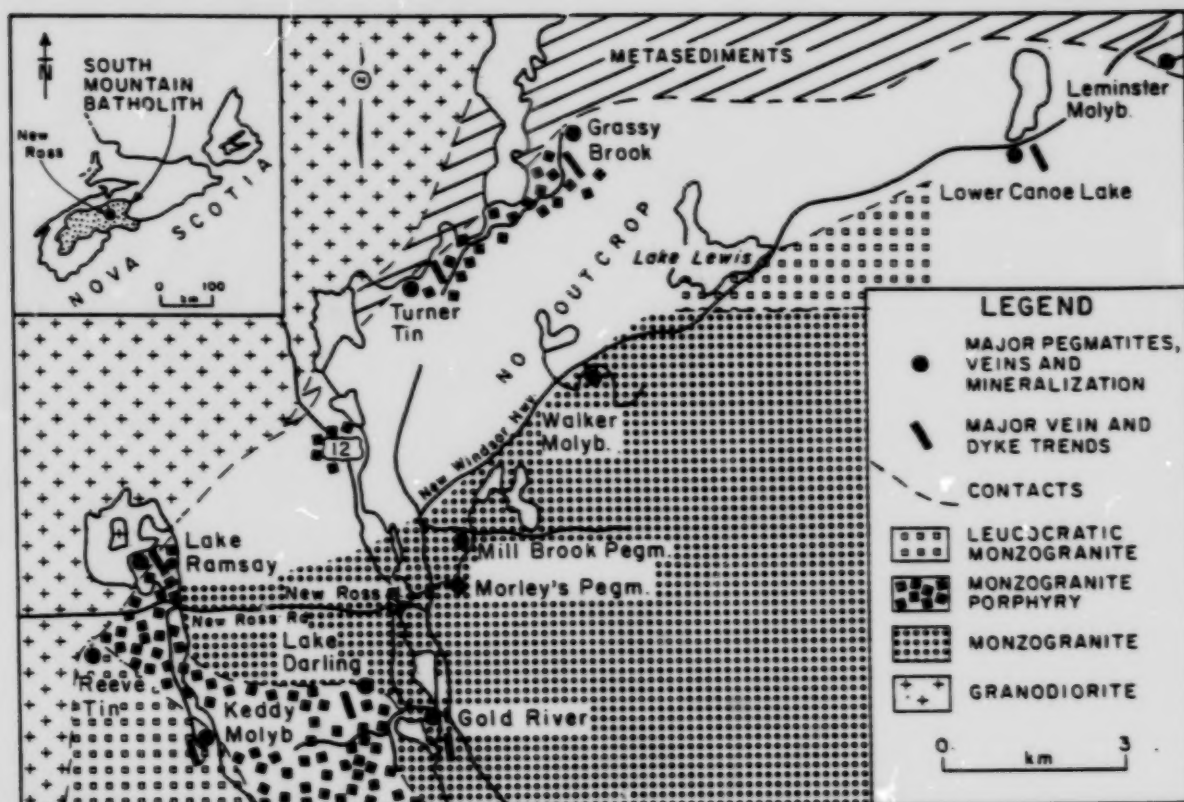


Figure 3-1. Geology of the New Ross area.

exponential, correlations amongst the elements are illustrated by Clarke, Muecke and Chatterjee (Fig. 1-2, this volume). Other correlations involving economically important trace elements, or economic indicator elements, will be shown here. A brief discussion of the geochemical behaviour of some of these elements follows.

Rubidium - As illustrated by Clarke, Muecke and Chatterjee (Fig. 1-2, this volume), Rb behaves as a typical large-ion lithophile element, and undergoes exponential increase in the later differentiates of the NRVC. When plotted against K_2O (or K), it can be seen (Fig. 3-2) that the range 200-300, considered by Shaw (1970) as the lower limit for magmatic processes. The trend in Figure 3-2a may be attributable to hydrothermal alteration in the NRVC.

Lithium - The principal silicate host for Li is mica (Heier and Billings, 1970) into which Li is partitioned (Kuts and Mischenki, 1963; Voronov and Lin, 1966) and thus magmatic differentiation processes involving extraction of phyllosilicates may result in a depletion of lithium. Despite this, Li is concentrated in later differentiates, especially in rocks which show signs of hydrothermal alteration. Unmineralized or unaltered granitoids contain about 40 to 100 ppm Li (Carron

and de la Roche, 1975; Groves, 1974; Kozlov et al., 1974; Heier and Billings, 1970).

Tin - Concentrations of tin increase with differentiation in mineralized rocks and is strongly correlated with lithium (Fig. 3-2b). Many of the rocks in the NRVC have $S_n > 20$ ppm, a concentration which is regarded as anomalous and indicative of ore deposits (Mulligan, 1975; Heep and Rigby, 1975). Mineralogical expressions of the high tin include cassiterite, stannite and wittichenite, all of which are associated with rocks involving hydrothermal fluids such as pegmatites and greisens. Micas also commonly concentrate tin (Barsukov, 1957).

Niobium-Tantalum - Members of the mossaite-tapiolite series are found as rare accessory minerals throughout the granites of the Meguma Zone. No increase in the enrichment of Nb is seen in the differentiation sequence although there is a 3.5 enrichment factor in Ta for the differentiation sequence. Similar observations have been made by others (Alexandrov and Laricheva, 1974). Even so, mossaite-tapiolite and columbite-tantalite are minor phases in some of the mineral deposits.

Table 3-1. Geology of the South Mountain Batholith: field relations and petrology

ROCK TYPE	FIELD RELATIONS	PETROLOGY			COMMENTS
		PRIMARY SILICATE ASSEMBLAGE	DEGREE OF ALTERATION	TEXTURES	
<u>QUARTZ VEINS</u>					
<u>GRANITE</u>	in gneiss; also in other phases generally as narrow (<1m wide) zones in host monzogranites or as late alteration in aplite pegmatite	Qtz ± Ksp	-	coarse grained associated with apilite-pegmatite; fine grained associated with quartz veins	may contain a large variety of rare minerals such as cassiterite, wolframite, molybdenite, amblygonite, fluorite and copper sulphides
<u>APLITE</u>	sometimes as continuous dykes in which either phase may dominate; more often as small irregular segregations	Qtz ± Ksp + Ms	minor in pegmatites; some apilites intensely greisenized	seccharoidal apilites and very coarse grained pegmatites	
<u>LEUCOGRAITIC</u>	small stocks or pods intrusive into monzogranites frequently causing strong alteration of host rock; sharp contacts observed	Qtz + Ksp + Pl(An ₃) + Ms ± Biot	moderate to strong	mostly medium but also coarse, porphyritic to hypidiomorphic granular (apilite endocontacta)	frequently associated with mineralized apilites, pegmatites and greisens
<u>MONOGRAITIC</u>	dykes a few to several metres wide with sharp contacts against monzogranite and monzogranite porphyry; associated pegmatites	Qtz + Ksp + Pl(An ₁₀) + Ms + Biot ± And	moderate to strong	Biot or Ms phenocrysts, or both, in saccharoidal groundmass	
<u>MONOGRAITIC</u>	irregular bodies; texturally gradational contacts observed with monzogranite	Qtz + Ksp + Pl(An ₁₀₋₃₀) + Ms + Biot ± And	moderate to strong	medium to coarse, porphyritic to hypidiomorphic granular	very few metasedimentary (Negama Group) xenoliths observed
<u>MONOGRAITIC</u>	small plutons to batholith size; sharp intrusive contact with granodiorite	Qtz + Ksp + Pl(An ₁₀₋₃₀) + Ms + Biot ± And	moderate to strong	coarse, porphyritic to hypidiomorphic granular	
<u>BIOTITE</u>	passively intruded into clastic sedimentary and volcanic rocks of Cambrian-Lower Devonian age	Qtz + Ksp + Pl(An ₃₀₋₄₀) + Biot	weak to moderate		abundant xenoliths at or near contact with metasediments

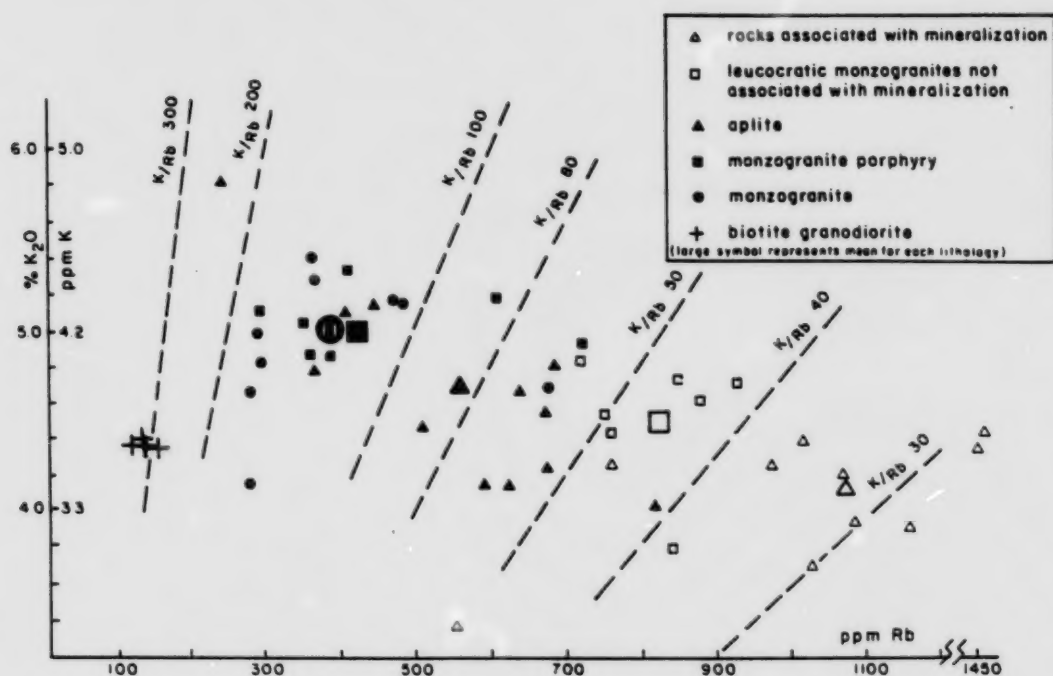


Figure 3-2a. K versus Rb.

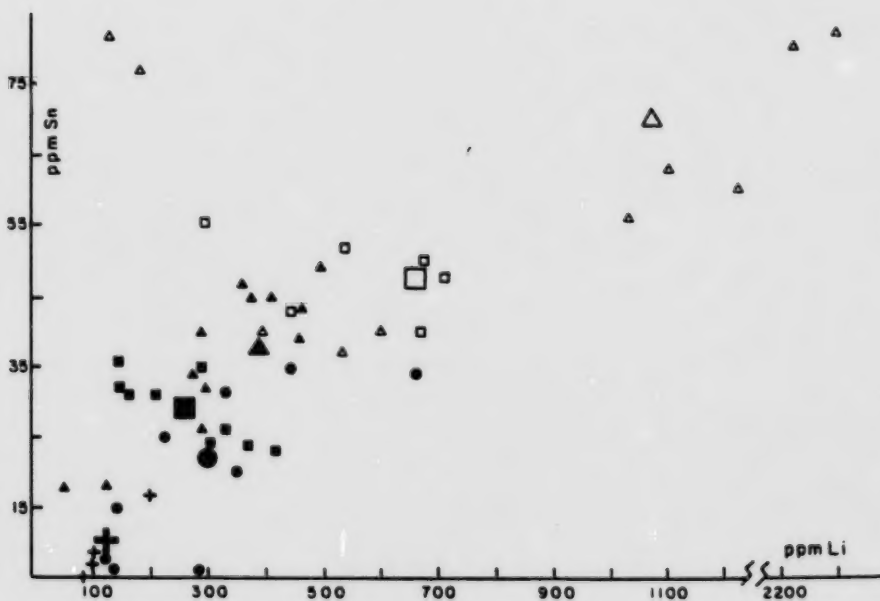


Figure 3-2b. Sn versus Li.

Copper - There is no systematic change in Cu concentrations through the differentiation sequence. Only in the mineral deposits themselves is evidence of very high Cu concentration observed in the form of a wide variety of primary and secondary copper sulphides and sulphosalts.

Molybdenum - Molybdenum similarly shows no marked change through the differentiation sequence and occurs as molybdenite in aplitic and pegmatitic rocks several deposits, including the Walker showing discussed later.

Tungsten - Most samples have concentrations close to the analytical detection limit of 1-2 ppm and analyses of host rocks gives no indication of the proximity to mineral deposits, at least one of which contains wolframite.

Zinc - Zinc shows a slight decrease through the igneous differentiation sequence and is found as minor sphalerite and gahnite in hydrothermal deposits.

Fluorine - Fluorine increases through the differentiation sequence, and the high activity of F is manifest in the occurrence of fluorite veins, topaz, durangite and the occurrences of biotite - andalusite through expansion of the stability field of biotite by fluorine. Muecke and Clarke (1981) have also alluded to the importance of F as a complexing agent for the REEs.

Phosphorus - Phosphorus shows a small increase through the differentiation sequence despite the apparent removal of significant amounts of apatite in the biotites of the granodiorites. Phosphorus reappears as an important element in the hydrothermal mineral deposits of U, as well as in fluorapatite, monazite and amblygonite in some other mineral deposits.

Of all the economically important elements concentrated in the mineral deposits, only Sn and Ta show a systematic increase through the fresh-rock of the igneous differentiation sequence and could be used to target new mineral deposits. The rest of the economically important elements are apparently not concentrated by an igneous process, and their high levels in the ore deposits are presumably due to the processes involving aqueous fluids, with or without complexing agents. The areas of mineral deposits are also characterized by high Li and Rb, and low K/Rb ratios as a result of hydrothermal fluid interaction and these characteristics can also be used to locate potentially important deposits.

MINERAL DEPOSITS OF THE NEW ROSS-VAUGHAN COMPLEX

Most known mineral deposits occur within the gravity low (Garland, 1953) centered on the town of New Ross. The principal showings are illustrated in Figure 3-1 and their field relations and mineralogy are summarized in Table 3-2. (For a more complete description, see O'Reilly et al., 1982). The deposits appear to be controlled either by host rock type, i.e. a close spatial relationship to the very late-stage leucocratic monzogranite, or by structure, i.e. contacts between the NRVC and the granodiorite, between

the NRVC and the metasedimentary country rocks, or along internal contacts within the NRVC. Both host rock type and structure may control the emplacement of a deposit.

Burnham and Ohmoto (1980) have summarized three types of processes that may lead to enrichment of economically important elements. The first of these is termed orthomagmatic which involves concentration of certain incompatible elements by a process of crystal-silicate liquid fractionation. The second is termed transitional which involves an aqueous fluid phase in association with, and usually derived from, a silicate magma with contained crystals. The aqueous fluid phase provides an important sink for the incompatible elements referred to above. The third process is simply termed hydrothermal and involves aqueous fluids and solid rocks at temperatures less than that of the water-saturated granite solidus. All three of these processes are believed to have operated in the South Mountain Batholith during cooling and examples are given below.

ORTHOMAGMATIC ENRICHMENT - THE LEUCOMONZOGANITES

It is clear from variation in modal mineralogy and from chemical variation diagrams (Fig. 1-2, Clarke, Muecke and Chatterjee, this volume; Fig. 3-2a, b) that the sequence granodiorite - monzogranite - porphyry - aplite - leucocratic monzogranite represents an igneous differentiation sequence. In that sequence some elements, such as Rb, Sn, and Ta show a progressive increase. The enrichment factors in the leucocratic monzogranites (relative to granodiorite) for these elements are: Rb-6.0; Sn-4.1; Ta-3.5. In no case, however, is it believed that orthomagmatic enrichment (modified by later hydrothermal processes) has led to any elemental concentration of economic interest. The value of orthomagmatic enrichment is that it may help to target (the so-called "metalotects" of Tschendorf et al., 1972) those plutons which may have mineral deposits built upon the primary or the magmatic concentration by the operation of fluid phases. Using this criterion, the leucocratic monzogranites are obvious areas for further exploration work.

TRANSITIONAL ENRICHMENT - THE WALKER POLYMETALLIC SN-W-CU-MO DEPOSIT

Perhaps the best example of transitional, or fluid-magmatic, enrichment leading to mineralization is to be found at the Walker deposit. The regional host rock is a leucocratic monzogranite and the deposit is apparently emplaced at or near a contact with monzogranite. Excavations have disrupted all surface exposure of the mineralized rocks, however early reports (Eardley-Willmot, 1925; Cooke, 1925), and dislodged samples around the workings, show it to be an aplite - pegmatite - greisen association. The aplite and pegmatite are intimately associated, clearly showing the coexistence of silicate magma and hydrothermal fluid. The same hydrothermal fluid was probably responsible for the associated greisenized rocks, including the aplite, pegmatite and others in which alteration is so

Table 3-2. Summary of mineral deposits data, New Ross-Vaughan Complex

NAME OF DEPOSIT	HOST ROCKS		STRUCTURAL CONTROL	RARE MINERALS	
	REGIONAL	LOCAL		OBSERVED	REPORTED
<u>LAKE RAMSAY</u>	monsoqr.	veins	contact		sphalerite, pyrite, chalcocopyrite, fluorite, Sn and Au minerals
<u>REEVES TIM</u> (Forties Prospect or King Pit)	leuco. monsoqr.	aplite pegmatite	contact	topiolite, cassiterite, lepidolite, amblygonite, apatite, fluorite	durangite, columbite, macrolite, tungstite, monazite, morinite, triploidite, triphylite, topaz, beryl, tourmaline, metatormerite, wolframite, scheelite
<u>KILBY MOUNT.</u> Lantz Prospect)	leuco.	aplite quartz veins	contact	molybdenite, topaz, fluorite	eosphorite, huebnerite, columbite (?), wolframite (?), fluorite
<u>LAKE CHARLIE AREA</u> (Larder River)	monsoqr. porph. to monsoqr.	veins (?) aplitic monsoqr.	none apparent	molybdenite, arsenopyrite (in greisenized host monsoqr. porph.)	Cu, Mo and Bi minerals
<u>TURNER TIM</u> (New Ross Prospect)	monsoqr. porph. to monsoqr.	veins greisen	contact	arsenopyrite, pyrite, sphalerite, stannite, chalcocopyrite, bornite, wittichenite, cassiterite, rutile, phenigite, covellite, chalcocite, fluorite, magnetite, wolframite	uranium assays
<u>CRASSY MINE</u> (Lake Wallaceback Tin)	monsoqr. porph. to monsoqr.	veins aplite pegmatite greisen	contact	cassiterite, arsenopyrite, pyrite, sphalerite, lepidolite, fluorite	eosphorite
<u>MORLEY'S PECK.</u> (Mill Stream)	monsoqr. porph. to monsoqr.	aplitic monsoqr. pegmatite quartz veins	none apparent	fluorite	Li, W, radioactive, "rare earths" minerals (old data)
<u>MILL BROOK PECK.</u>	monsoqr.	pegmatite aplite	none apparent	fluorite	lithium mica
<u>WALKER MOUNT.</u> (New Russell Prospect)	leuco. monsoqr.	aplite pegmatite greisen	contact (?)	<u>Aplite/Greisen:</u> molybdenite, cassiterite, huebnerite, bornite, chalcocopyrite <u>Pegmatite:</u> molybdenite <u>Greisen:</u> Wolframite, gahnite, digenite, bornite, covellite, uraninite(?), Mn-rutile, chalcocopyrite, tennantite, wittichenite, native Bi <u>Also:</u> phenigite, siderophyllite, fluorite, tourmaline, chalcocite, topaz	
<u>LOWER CANOE LAKE</u>	monsoqr.	aplitic monsoqr. veins little pegmatite	none apparent	fluorite, tourmaline	
<u>LEMINSTER</u> (Swimmer Prospect)	metasediment	veins	contact (?)		molybdenite, arsenopyrite, specularite
<u>HAYVERTILL</u> (Russell Pit)	monsoqr. (?)	pegmatite	(?)		tourmaline, Sn-minerals
<u>SEFTERVILLE</u> (Gold Mine Hill)	monsoqr.	pegmatite aplite (?)	(?)		lepidolite, tourmaline, fluorite

complete that the protolith is unidentifiable. The aplite and pegmatite consist of an assemblage of quartz, K-feldspar, plagioclase and muscovite whereas the greisen consists of phengite, siderophyllite, quartz, topaz, tourmaline and fluorite.

The following associations are observed: molybdenite is associated with the aplite and pegmatite, and oxides (wolframite, gehnite, cassiterite, uraninite) and other sulphides (bornite, mawsonite, tennantite, wittichenite, chalcopyrite) are largely associated with greisenized rocks. Evidence for the early development of oxides includes mawsonite flakes which lie in cracks in cassiterite, sulphides which partially enclose wolframite and gehnite, and covellite (altered bornite?) which is found in fractures in wolframite (Fig. 3-3a). Both wolframite (Hsu, 1976) and cassiterite (Moh, 1974) can coexist with sulphides; however the presence of mawsonite in cracks in cassiterite indicates that, in the later stages of deposition, tin was incorporated into Cu - Sn sulphides rather than oxides. Bismuth metal occurs in blebs and disseminations in mawsonite, wittichenite and tennantite (Fig. 3-3b).

Molybdenite is probably the highest temperature mineral in the deposit. It is stable above 610° C (Stemprok, 1966) and can be taken as a rough estimate of the upper temperature limit of deposition. The association bornite-mawsonite is only stable below 390° C (Lee et al., 1975). Neither mawsonite nor wittichenite is associated with chalcopyrite; this indicates the presence of a bornite solid solution (> 500° C, Sugaki et al., 1975) zone which has a tie line with wittichenite. Chalcopyrite would therefore appear to be a late phase. Native bismuth has a precipitation temperature of 271° C, placing a lower temperature limit on deposition at the Walker deposit.

The molybdenite is believed to have precipitated from the pegmatitic fluid in equilibrium with a silicate melt represented by the aplite. The other opaque minerals at the Walker deposit were precipitated from greisenizing fluids, at least in part emanating from the pegmatitic residue and affecting the surrounding rocks. The sequence of phases precipitated by this fluid was dependent on temperature and fugacity conditions (fO_2 , fS_2) in the cooling assemblage and, judging from the temperature range of deposition, the deposit may have formed over a long period of time.

HYDROTHERMAL ENRICHMENT - THE TURNER TIN DEPOSIT

The Turner deposit was first described by Faribault (1908, 1910, 1911) and subsequently by Douglas and Campbell (1942) and Cameron (1950). It lies within 100 m of the large inlier of meta-sedimentary rocks of the Meguma Group in a host of monzogranite which is pink in colour and has undergone extensive alteration. The mineralization occurs in three quartz vein-greisen zones and one quartz porphyry. A summary of the principal occurrence, the Discovery Vein, is given in Table 3-3, and further details of the

others may be found in Farley (1979).

The following description of the opaque mineralogy is based largely on material from the Discovery Vein which is best exposed. The opaque minerals were deposited during greisenization and are almost invariably found in patches of sericite and phengite, or enclosed in nearby quartz, or associated with veins of iron-rich material now altered to goethite or limonite. The textural relations are as follows: arsenopyrite is found surrounded by both pyrite (Fig. 3-4a) and chalcopyrite; pyrite occurs in cubes and skeletal masses closely intergrown with K-feldspar and phengite.

Sphalerite, with inclusions of pyrite, occurs mainly disseminated in the greisenized central quartz vein. It is closely associated with chalcopyrite which frequently completely envelops the sphalerite. Some of these latter sphalerite grains also contain minute inclusions of chalcopyrite, which may be the result of exsolution or of contemporaneous deposition of the two minerals. Sphalerite has been extensively replaced by covellite, sometimes in a zonal pattern (Fig. 3-4b). Even where sphalerite is enclosed in chalcopyrite, it is generally more altered than the latter.

Chalcopyrite is disseminated in the greisen and the central quartz vein where it is in direct contact with all the other opaque phases. It envelops arsenopyrite (partially), sphalerite, pyrite, stannite, and is intergrown with wittichenite and bornite. Bornite occurs in small patches in chalcopyrite and contains exsolved chalcopyrite lamellae (Fig. 3-4c).

Stannite is rare and is mainly enclosed in chalcopyrite (Fig. 3-4d) although one isolated grain was found. Wittichenite occurs in a similar association, being enclosed by or closely intergrown with chalcopyrite, but some grains occur independently in gangue.

Cassiterite is found in grains of size ranging from 0.3 mm to 4 mm, with an average of 0.5 mm. It frequently occurs closely intergrown with chalcopyrite.

Following emplacement of these opaque minerals, the vein and greisen zone were subjected to minor deformation as evidenced by undulatory extinction and granulation of earlier grains. A second generation of quartz veins cut the greisen and central quartz vein and these veins may have central bands of K-feldspar or may be cut by K-feldspar veins. These feldspars were then partially kaolinized by very late-stage deuteric alteration.

It is proposed on the basis of the relations described above that water-rich hydrothermal fluids were injected along a fracture in the consolidated roof zone of the granite. The location of the fracture may have been controlled by the proximity of the large Meguma inlier. In the case of the quartz vein greisen zones there is no evidence of any contemporary silicate magma, although a quartz porphyry, barren except for some fluorapatite, is associated with minor

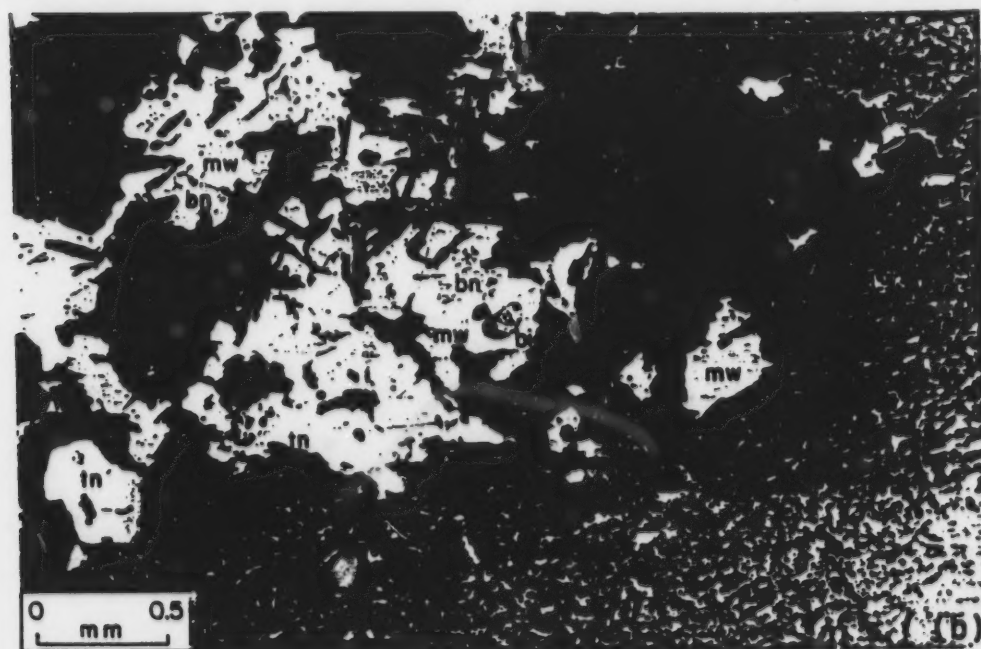


Figure 3-3. Mineral paragenesis in the Walker deposit. (a) This sample (W4) shows a large grain of bornite enclosing mawsonite and intergrown with tennantite and wittichenite. Covellite is replacing bornite along the cracks. Reflected light. (mw - mawsonite, wt - wittichenite, tn - tennantite, bn - bornite, cov - covellite).

(b) A bleb of bismuth metal lies in mawsonite. The mawsonite is closely intergrown with tennantite and bornite. Reflected light. (bi - native bismuth).

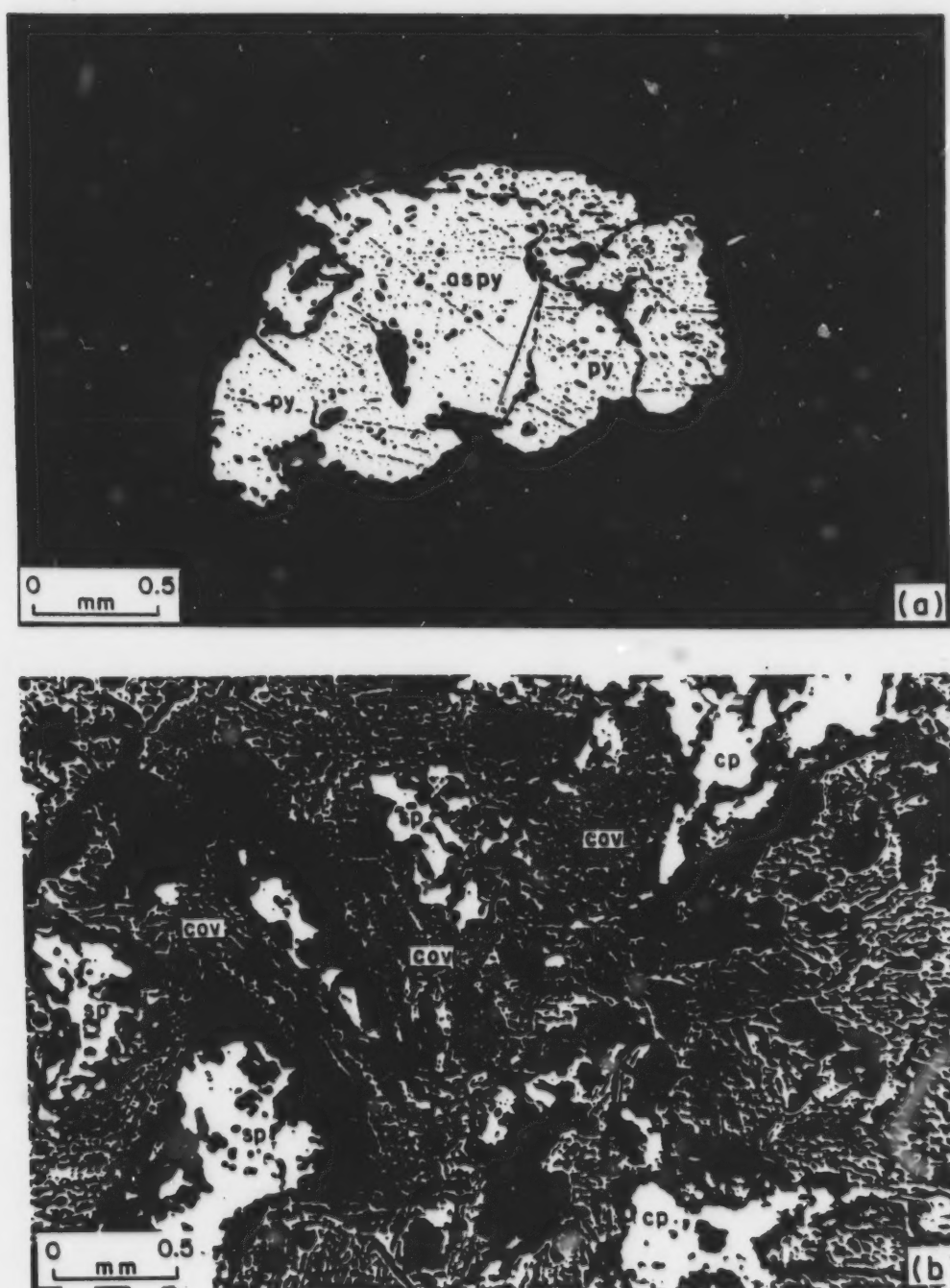
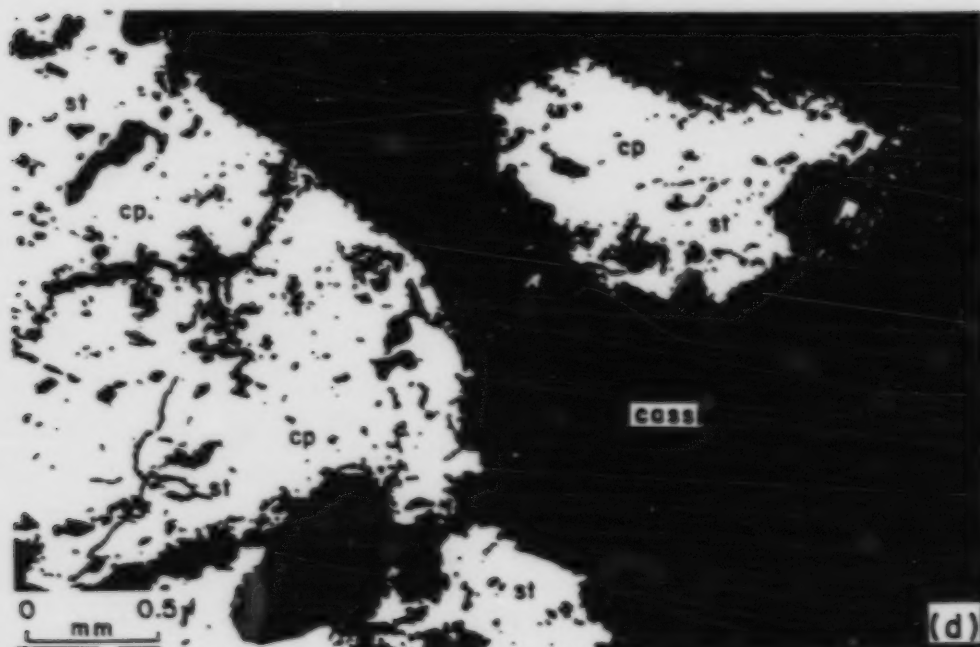
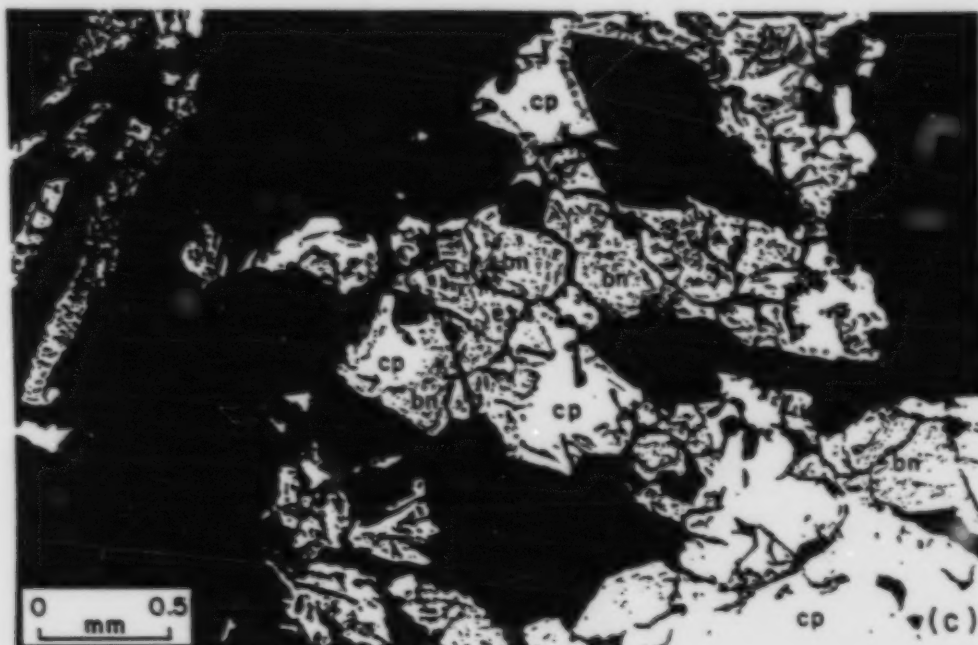


Figure 3-4. Mineral paragenesis in the Discovery Vein at the Turner deposit. (a) Arsenopyrite is enveloped by pyrite. Reflected light. (aspy - arsenopyrite, py - pyrite). (b) Sphalerite is being zonally replaced by covellite; chalcopyrite is largely unaltered. Reflected light. (sp - sphalerite, cov - covellite, cp - chalcopyrite).



(c) Bornite is shown closely intergrown with chalcopyrite, surrounding clear patches of chalcopyrite and in turn enveloped by chalcopyrite. Bornite is exsolving chalcopyrite (exs.) Reflected light. (bn - bornite, cp - chalcopyrite). (d) This sample shows stannite enclosed in chalcopyrite intergrown with cassiterite. Reflected light. (cp - chalcopyrite, st - stannite, cass - cassiterite).

Table 3-3. Summary of petrological data, discovery vein, Turner deposit

NIPT ROCK	TRANSITION OF SECTION	GREENS	VEINS
Pink porphyritic monzonite. Coarse grained (10-5 cm). Orthoclase laths up to 2 cm. Red hematite (some) plagioclase up to 7 mm. Aggregates of quartz up to 7 mm across. Muscovite and biotite.	Over 2 cm. Chloritized biotite. Sericitized plagioclase (hematite overgrowth). Sericitized and muscovitized K-feldspar. Quartz - primary and secondary. Muscovite - primary and secondary.	Green. Quartz-rich with much sericite and chlorite.	Quartz. Crystal 0.2 x wide semi-circular throughout length of trench. Quartz stringers cut the greisen and quartz vein (1.5 cm x 2 cm). Microclitic quartz. K-feldspar veins 1 cm x 1 cm cut the quartz stringers and greisen.

greisen and may represent the very last magma to reach this level in the carapace. The most likely source of hydrothermal fluid was the vast thickness of crystallizing, water-saturated granite magma lying beneath the present level of erosion in New Ross. Interaction with fluids of meteoric origin, however, is also possible since the SMB was emplaced in the epizone (McKenzie and Clarke, 1975).

A MODEL FOR THE EVOLUTION OF THE CENTRAL PART OF THE BATHOLITH

The three-dimensional shape of the SMB is similar to that of a mushroom with its stalk extending to a depth of 25 km in the vicinity of New Ross (Garland, 1953; Douma, 1978). Elementary cooling models suggest that this region of thickest granite may have been both the last place in the batholith to experience magmatic activity and the last place in which the magmatic thermal anomaly decayed away. It is clear from Table 1-1 of Clarke, Muecke and Chatterjee (this volume) that the most evolved differentiates of the Batholith occur in the New Ross area, and Rb-Sr radiometric age dating (Clarke and Halliday, 1980) has confirmed that, indeed, the youngest magmatic rocks exist in this region. If the stalk of the mushroom represents a region of diapiric ascent of magma, then New Ross is clearly the magmatic focal point of the Batholith, and many diapirs of varying sizes, and at different times, may have used this conduit. We believe that the SMB has evolved in a manner similar to Ivulume batholith, as described by Bateman and Chappell (1979), wherein earlier solidified or semi-solidified parts of the Batholith have been diapirically penetrated by their later comagmatic differentiates (Fig. 3-5). This process continued as long as the buoyant forces in deeper gravitationally unstable magmas were greater than the strengths of the overlying roof rocks.

Some of the larger diapiric upwellings may have completely penetrated the granodiorite envelope to come in direct contact with the Meguma country rock, perhaps as is suggested by the location of the Springfield Lake monzogranite body (McKenzie, 1974) and, indeed, the NIPT itself if the large inlier of Meguma meta-sedimentary rocks is a roof pendant. If contamination of the granite magma by meta-sediments is required to explain the chemical evolution of the Batholith, then this mechanism

permits continued assimilation of new material after the granodiorite stage of crystallization. Some of the smaller diapirs probably never reached the exocontact and are now exposed as irregular, and often gradational, bodies of monzogranite and porphyry in the New Ross area. Discrete bodies of leucomonzogranite, often with fine-grained margins and associated alteration haloes in the host rocks, appear to have been the latest bodies emplaced diapirically. Thereafter, subvertical joints in the earlier bodies served as avenues for barren mica splite dykes, apatite-pegmatite associations such as the Walker deposit, and hydrothermal veins such as in the Turner deposit (Fig. 3-6). The commonly-observed close spatial association of leucomonzogranites and mineral deposits may be genetically significant in that the leucomonzogranites may be the source of the fluids.

No systematic study of the ages of the mineral deposits has yet been done, but some may be as much as 70 million years younger than the last magmatic deposits. In these cases, it is not clear yet whether two distinct episodes are involved, i.e. unrelated magmatic and hydrothermal events or, more likely, as demonstrated by the Walker deposit, a continuum of deposition from the magmatic to the low temperature hydrothermal minerals. Although the source of the early waters and economic elements is almost certainly magmatic, it is not clear whether all later waters and metals were from the same magmatic source. This is one of the principal questions to be answered by further work.

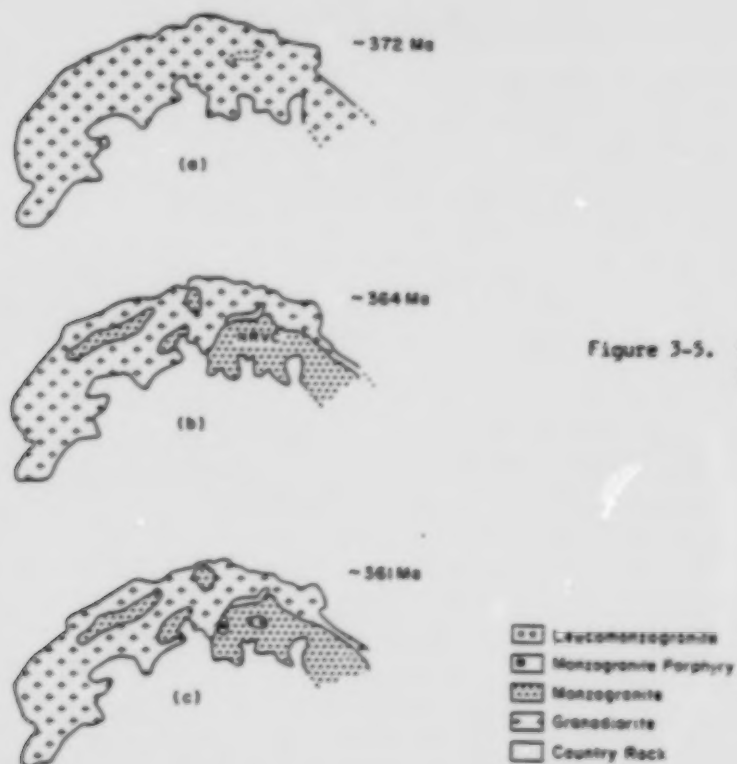


Figure 3-5. Sequential intrusive relationships in the South Mountain Batholith.

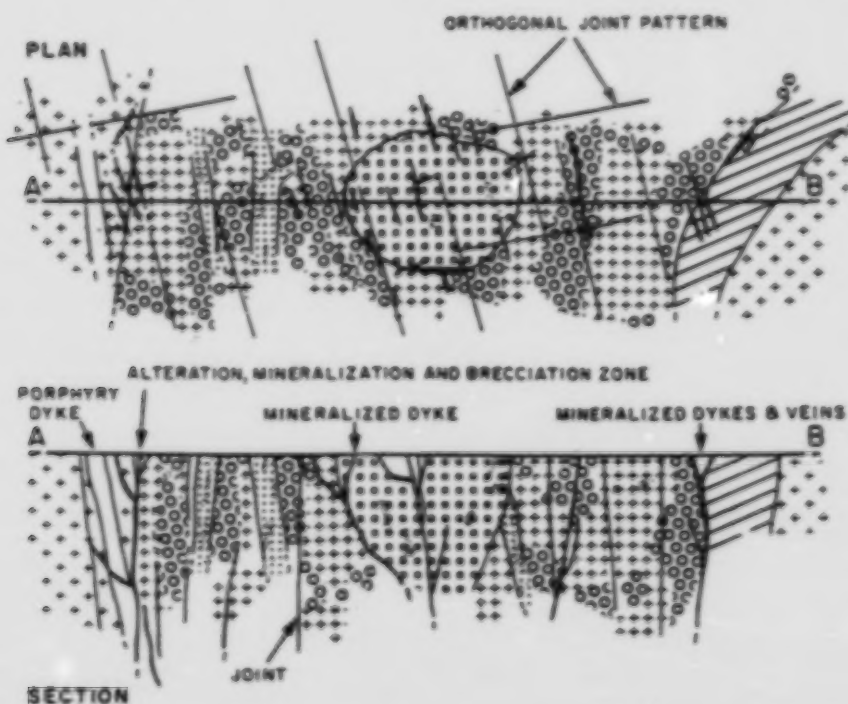


Figure 3-6. Schematic illustration of evolution model.

CHAPTER 4

ECONOMIC GEOLOGY OF THE NEW ROSS-VAUGHAN COMPLEX

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Logothetis, J., Economic geology of the New Ross-Vaughan Complex; in Guide to the Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

INTRODUCTION

The granitoid rocks comprising the South Mountain Batholith (SMB) are host to many occurrences of Sn, W, Mo, U, Zn, Cu and Mn mineralization. With the exception of the manganese deposits near New Ross, none of these occurrences has yet been mined. In recent years, however, mineral exploration in the SMB has resulted in the discoveries of deposits of tin near East Kemptville (presently under mine development) and uranium at Millet Brook. These successes have done much to confirm the long-suspected potential of the SMB with respect to economic mineralization of Sn-U-Mo-Cu-Zn.

This contribution reviews the principal types of mineralization recognized in the Batholith with greater emphasis placed on the mineral occurrences in the region of the New Ross-Vaughan Complex (NRVC). In addition, the mineralogical, geochemical and geological features of the SMB are examined with regard to finding useful indicators that could distinguish granitoid bodies which have a high potential for tin mineralization from those having a lower potential. Towards this end, the granitoid rocks of the SMB are compared with other well-documented granitoid complexes which contain tin deposits.

MINERALIZATION IN THE NEW ROSS-VAUGHAN COMPLEX

Recent compilation maps of the metallic mineral occurrences in the SMB (Gregory et al., 1979; Chatterjee, 1983) indicate the presence of a large number of mineral showings (e.g. principal types of mineralization include Sn, U, W, Mo, Cu, Zn, Mn, Pb, Ag and Li). Many of these occurrences are polymetallic with Sn-W-Zn-Cu, Sn-U and U-Cu being some of the commonly recognized associations (e.g. Chatterjee, 1983).

A large proportion of the known mineral occurrences in the SMB are concentrated in the vicinity of the NRVC (Fig. 4-1). The geophysical features of this region, i.e. the presence of a large negative gravity anomaly and high heat flow values (Garland, 1953; O'Reilly, 1975; Rankin, 1974) suggest that the NRVC is the thickest part of the Batholith. The presumed greater volume of magma in this area is considered to have been favourable for the slow cooling of granitic melts, and by implication, the evolution of late stage fluids enriched in incompatible elements (Sn-W-Mo-Cu-U) responsible for mineralization (Farley, 1979; Muecke and Clarke, 1981).

Most of these mineral occurrences were first discovered and worked by the local residents during the turn of this century. They have been described in a number of reports by the Geological Survey of Canada and Nova Scotia Department of Mines (e.g. Hanson, 1932; Douglas and Campbell, 1942a, 1942b; Weeks, 1946; Ellsworth, 1932; Little, 1959; Vidnes, 1960; Mulligan, 1975). Comprehensive field-mineralogical-geochemical investigations employing modern techniques have been carried out by Bishop and Wright (1974), Charest (1976), Felderhof (1978), Farley (1979), O'Reilly et al. (1982), and

Chatterjee et al. (1982) on different aspects of mineralization.

The better known metallic mineral occurrences in the NRVC are listed in Table 4-1 together with other important occurrences from elsewhere in the Batholith (see Figure 4-1 for locations). The following account of mineralization in the Batholith is based largely on occurrences found within the NRVC. Four principal types of mineralization phenomena can be recognized:

- (i) polymetallic tin mineralization
- (ii) molybdenum
- (iii) uranium
- (iv) manganese
- (v) other types of mineralization

POLYMETALLIC TIN MINERALIZATION

Polymetallic tin mineralization (i.e. $\text{Sn} \pm \text{W} \pm \text{U} \pm \text{Mo} \pm \text{Cu} \pm \text{Zn} \pm \text{Bi} \pm \text{Ag} \pm \text{Li}$) is almost exclusively associated with greisens and greisenized granitoid rocks (localities 1-8, Table 4-1). This type of mineralization is spatially and genetically related to the more differentiated biotite-muscovite monzogranites and muscovitic ($\pm$ biotite) leucomonzogranite bodies which have undergone late stage greisen-type alteration (Table 4-1). The mineralized greisen bodies which are commonly associated with the roof, or border zones (or both) of the NRVC (e.g. Charest, 1976; O'Reilly et al., 1982) occur in two distinct modes:

- (1) as selvages localized along the flanks of steeply-dipping quartz $\pm$ sulphide veins (e.g. quartz-sericite greisen (QS), quartz-sericite-pyrite-fluorite greisen (QSPF), Turner Tin, Lake Darling and East Kemptville), quartz porphyry dyke bodies (e.g. elvan veins, Turner Tin prospect), pegmatitic veins (Walker prospect, Farley, 1979), and
- (2) as irregularly shaped zones of pervasive greisenization typically associated with muscovitic leucomonzogranite bodies, not spatially related to veins or fractures (or both) (e.g. Long Lake, Wallaback Lake and Keddy prospects, Table 4-1).

Mineralization, which is complex and highly variable (Table 4-2a), occurs as vein infill, as disseminated minerals deposited in open spaces, and as replacement of pre-existing minerals. Cassiterite, chalcocite, chalcopyrite, bornite, sphalerite, wolframite, scheelite and arsenopyrite are commonly identified. Less frequently recognized species may include tennantite (Sb), wittichenite (Bi) and Bi-metal, gahnite (Zn), huebnerite (W), uraninite and torbernite (U), and columbite (Nb,Ta) (many references: e.g. Charest, 1976; Farley, 1979; O'Reilly et al., 1982; see Table 4-2a).

Greisenization of the host granitoid rocks typically gives rise to quartz-phengite (greisen) assemblages containing variable amounts of chlorite, rutile, albite, K-feldspar, phosphate

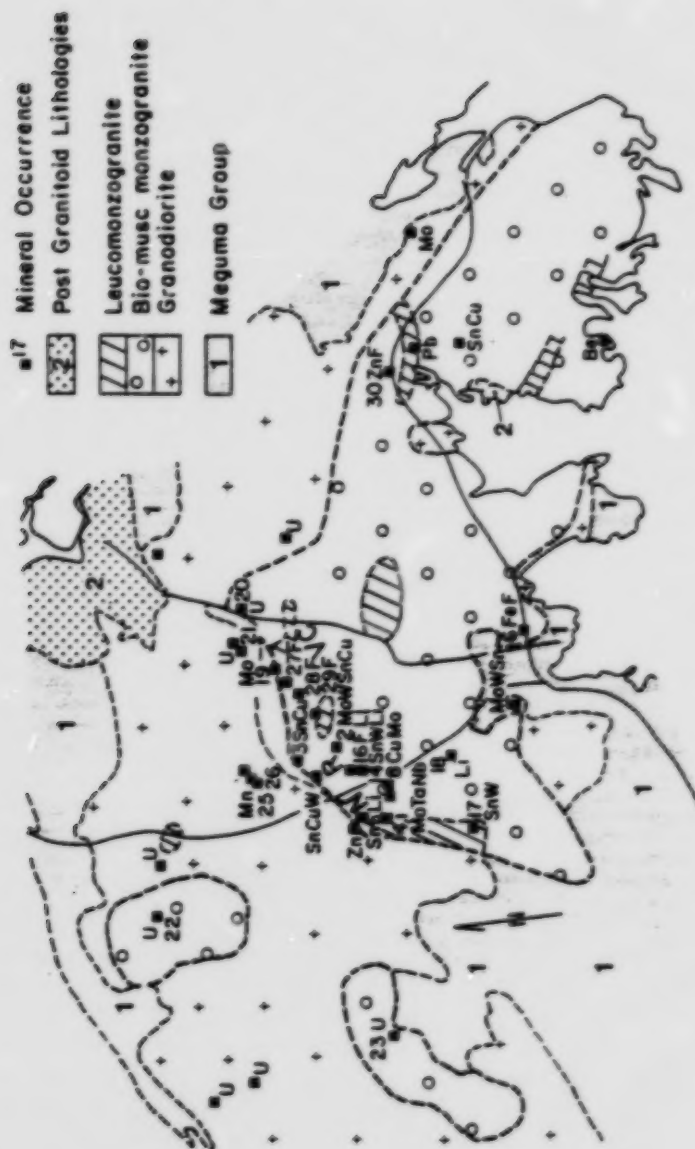


Figure 4-1. Geological map of southern Nova Scotia showing the eastern half of the South Mountain Batholith (SMB) and distribution of granitoid rock types. The location of most of the well known mineral occurrences and prospects in the New Ross-Vaughan Complex (NRVC) are shown (numbering of localities same as Table 4-1).

Table 4-1. Metallic mineral occurrences and prospects of the New Ross-Vaughan Complex and from elsewhere in the Batholith. The occurrences are listed under the headings of the principal types of mineralization discussed in text. The numbers associated with each occurrence may be used to locate its position in Figure 4-1.

Polymetallic tin mineralization:		
1. Long Lake	W, Sn, Pb, Cu, Mo, Ag, (Rb, Cs, Li, REE)	greisens (irregular pods and veins), greisenized leucomonzogranite and Mo-bearing pegmatite
2. Walker Moly	Mo, W, Sn, Cu, Bi, Sb, Zn, U	massive greisens, greisenized leucomonzogranite host, and Mo-bearing pegmatite
3. Walleback Lake	Sn, Zn, Li	irregular greisen patches, greisenized leucomonzogranite, and pegmatite
4. Keddy	Ta, Nb, Mo	pervasively greisenized leucomonzogranite, and Mo-bearing pegmatite
5. East Kemptville	Sn, Cu, Zn	greisenized leucomonzogranite and greisen-bordered quartz veins
6. QSPV	Fe, Cu, fluorite	greisen-bordered quartz veins in biotite-muscovite monzogranite
7. Turner Tin	Sn, Zn, Cu, Fe, Bi	greisen-bordered quartz veins in biotite-muscovite monzogranite, and associated silver veins with greisenized borders
8. Lake Darling	Cu, (Mo)	Mo-bearing quartz-feldspar vein (pegmatite?) with greisen selvages in biotite-muscovite monzogranite
Molybdenum:		
9. Long Lake	Mo	Molybdenite-rich metasomatic "pegmatite"
10. Walker Moly	Mo	Molybdenite-rich pegmatite
11. Swinimer occurrence	Mo	quartz veins in Meguma Group
12. Keddy	Mo	Molybdenite-bearing pegmatite and leucomonzogranite
13. Lake Darling	Mo	Molybdenite-bearing quartz vein
Pegmatite-hosted occurrences:		
14. Morley's Pegmatite	Sn, W, Li, F	pegmatite dykes and associated breccia (jasper breccia) in leucomonzogranite and monzogranite
15. Reeves Pegmatite	Sn, Li, F	pegmatite in leucomonzogranite
16. Mill Brook	F, Li	pegmatite dyke in biotite-muscovite monzogranite
17. Nevertell Lake	Sn, W	pegmatite dyke in leucomonzogranite
18. Sefterville pegmatite	Li	quartz and pegmatite dykes in monzogranite
Uranium:		
19. Vaughan Episyenites	U, Zn, (P)	mineralized episyenite pods or lenses in biotite-muscovite monzogranite developed next to fine grained (porphyritic) monzogranite dykes
20. Millet Brook	U	uraniferous shear/fracture zones in granodiorite
21. Casperau Lake	U	uraniferous shear/fracture zones in leucomonzogranite
22. East Dalhousie	U	autunite bearing fractures and/or shear zones in altered biotite-muscovite monzogranite
23. Many occurrences (e.g. Turner Tin)	U	associated with greisen hosted polymetallic tin mineralization
Manganese:		
24. Cain-Riddle Mine		manganiferous shear/fracture zones in granodiorite
25. Dean and Chapter Mine		
Fluorite & sulphide veins:		
26. Lemster	F	fluorite veins
27. Cane Lake	F	fluorite veins
28. Lake Lewis	F	fluorite veins
29. Upper Tantallon	Zn, F	fluorite & sphalerite veins

minerals (e.g. apatite and triplite), and F- or B-rich (or both) phases (e.g. topaz, fluorite, and tourmaline) (Table 4-2b). Detailed investigations of greisen bodies reveal a complex paragenetic history of secondary mineral formation (e.g. Farley, 1979; Richardson et al., 1982). At QSPF and Long Lake occurrences the effects of various alteration types (e.g. chloritization, rutilization, silicification, fluoritization, pyritization, and K-feldspathization) are superimposed on the pre-existing greisen minerals producing spatially zoned alteration assemblages. Zoned alteration envelopes flanking thin quartz veins have been described by Richardson et al. (1982) from East Kemptville deposit. The greisens in this locality are characterized by an inner topaz-sulphide zone, which with increasing distance from the veins grades into a quartz-topaz-muscovite-sulphide zone, a green mica-quartz zone and an outer sericitized granite.

MOLYBDENUM

Molybdenum mineralization occurring as molybdenite-bearing pegmatites (e.g. Long Lake, Walker Moly, Keddy prospects), and within greisenized leucomonzogranites (e.g. Keddy prospect) and quartz veins (e.g. Swinimer's occurrence) has been described from the NRVC (e.g. Charest, 1976; Farley, 1979; O'Reilly et al., 1982; see Table 4-1). In general, molybdenum mineralization is spatially associated with leucomonzogranite and greisenized leucomonzogranite bodies. Sharp intrusive contacts between the pegmatites and the surrounding leucomonzogranites are recognized at the Wallaback Lake and Keddy prospects and at the leucomonzogranite/Meguma contact at Long Lake. Fragments of the enclosing leucomonzogranite are reported to occur within the pegmatite at the Walker Moly prospect (Farley, 1979). Together, these observations indicate that the pegmatites are later and have intruded the leucomonzogranite host bodies.

The origin of molybdenum mineralization is open to two different interpretations depending on whether the molybdenite-bearing pegmatites are considered to be derived by dominantly:

- (1) "magmatic and transitional processes" (i.e. as products of end-stage magmatic differentiation); or
- (2) "hydrothermal processes" (i.e. related to the metasomatic alteration of the enclosing granitoid rock).

Both views regard the pegmatites as late-magmatic phenomena.

The first model (i.e. the "magmatic-transitional model") has been proposed by Farley (1979) to account for molybdenum and accompanying polymetallic tin mineralization at the Walker Moly prospect. She suggested the intrusion of the molybdenite-rich pegmatite resulted in the formation of the mineralized greisens, by the interaction of mineralizing fluids from the pegmatite with the enclosing leucomonzogranite host. According to this view, molybdenum was preferentially deposited in the pegmatite, whereas, the other metallic elements (Sn-W-Cu-Zn-Bi) were

therefore, greisenization follows, and is controlled by, the intrusion of a metal and sulphur-rich pegmatitic vein.

Alternatively, the second view (i.e. the "metasomatic model") is proposed to explain molybdenum mineralization associated with the Long Lake prospect. In this model, the origin of the "pegmatite-looking", quartz + K-feldspar veins and accompanying molybdenite mineralization are considered to be related to:

- (1) the large-scale metasomatism of Si, Ca, Na and K associated with the greisenization of the host leucomonzogranite; and
- (2) the high solubility of molybdenite in alkaline, high-temperature hydrothermal fluids.

With regard to the first observation, the redistribution and subsequent deposition of Si, Ca, Na and K leached during greisenization may give rise to the formation of metasomatic "pegmatite" bodies. This view is supported by mass-transfer calculations which indicate that greisenization of one kg of leucomonzogranite was accompanied by the removal of up to 370 g of SiO_2 , 10 g of K_2O , 30 g of Na_2O , and 5 g of CaO . In view of the large volume of rock affected by greisenization, deposition of only a small fraction of these leached constituents would suffice to produce the pegmatitic veins at Long Lake. Fractures within the leucomonzogranite/Meguma contact, and possibly the enclosing Meguma wall rock, appear to have been suitable sites for their redeposition.

With regard to the solubility of molybdenite, Isak and Carman (1976) have experimentally shown that molybdenite is remarkably soluble at high temperatures in the system $\text{Na}_2\text{Si}_2\text{O}_5 - \text{MoS}_2 - \text{H}_2\text{O}$ (i.e. up to 12.5 wt.% MoS_2 at 650°C, and 1.4 kb). In the light of the above comments, the mobility and solubility of molybdenum and sulphur were presumably enhanced in the alkali-bearing greisenizing fluids at Long Lake. Hence, the formation of metasomatic pegmatite veins at Long Lake and accompanying molybdenite mineralization are regarded as genetically interrelated and produced as a consequence of greisenization.

Although two alternative models are here considered here, they need not be regarded as mutually exclusive and competing explanations of molybdenum mineralization. Rather, it is more

likely that both models need to be evoked to adequately account for the complete spectrum of molybdenum mineralization found in the NRVC.

URANIUM

Although the list of well-studied uranium occurrences in the SNB are limited in number, it is convenient to distinguish on the basis of their geological and structural features three different types of uranium mineralization.

Greisen-Hosted Uranium Mineralization

Since greisen-hosted polymetallic mineralization (with/without uranium) has already been

discussed, this type of uranium association is only briefly described below. Greisen-hosted uranium mineralization is known from a number of occurrences within the NRVC, e.g. Walker Moly and Turner Tin (e.g. Farley, 1979; O'Reilly et al., 1982), and from elsewhere in the Batholith (e.g. Cloud Lake and Paradise Lake: from Mineral Occurrence Data Cards, Nova Scotia Department of Mines). Uranium phosphates, chiefly torbernite or autunite (or both), and uraninite (Walker prospect) have been identified in these occurrences. At Long Lake, uranium shows a strong enrichment in the greisen zones (up to 76 ppm compared to only 18 ppm for the host rock). Independent uranium-minerals, however, were not identified by the author.

Uraniferous Shear and/or Fracture Zones

Under this heading the main features of three important uranium occurrences are briefly described, i.e. the Millet Brook deposit, and the East Dalhousie and Gaspereau Lake prospects. All three occurrences are located outside the NRVC proper (see Fig. 4-1 for locations). Although these occurrences are associated with shear/fracture zones in altered granitoid host bodies, and hence are grouped together for discussion, a common origin with respect to uranium mineralization is not implied.

Uranium mineralization is known to occur as uraniferous shear or fracture (or both) zones within host granodiorites near Millet Brook (Chatterjee et al., 1982), biotite-muscovite monzogranites in the vicinity of East Dalhousie and leucomonzogranites at the Gaspereau Lake occurrence (O'Reilly et al., 1982). Mineralization, though highly variable (Table 4-2a), is dominated by minerals of essentially uranium phosphate composition (e.g. autunite, torbernite, meta-torbernite) often with associated pitchblende. The effects of hydrothermal alteration on the mineralogical and geochemical composition of the host granitoid rocks in close proximity to the mineralized shear/fracture zones have been described by O'Reilly et al. (1982), and Chatterjee et al. (1982) (see list of secondary minerals in Table 4-2). Since the Millet Brook deposit has been described in some detail by Chatterjee et al. (1982) and Chatterjee (this volume) its main features deserve little further comment.

Mineralization at Millet Brook occurs as closely spaced en echelon pods and lenses within northeast-trending shear or fracture zones that cut the biotite granodiorite. Mineralization is dominated by pitchblende, torbernite, Pb-meta-autunite with minor amounts sulphides (e.g. chalcopyrite, covellite and galena). Intense alteration of the host granodiorite may extend up to 30 metres from the mineralized fractures and includes the following types: albitization, muscovitization, biotitization, chloritization, K-feldspathization and carbonitization. Chatterjee et al. (1982) proposed that uranium mineralization at Millet Brook postdates the intrusion of the granodiorite, and is probably related to late stage felsic magmatism of a proximal, yet unexposed, leucomonzogranite intrusive body.

Episyenite Zones

The Vaughan uranium occurrence (mineral deposit 20 of Fig. 4-1 and Table 4-1) is associated with metasomatized granitoid rocks of episyenite composition, i.e. essentially albite + K-feldspar + chlorite + phengite assemblages. This type of uranium association has not been previously documented in the SMB.

Uranium mineralization at Vaughan is confined to episyenite zones situated within the roof zone of the NRVC. These episyenite zones occur as sheets/lenses mainly within a medium grained monzogranite host which has undergone intense episyenitization along the flanks of a number of fine grained porphyry dykes. The dykes, ranging in size from several centimetres to over 12 metres in width, are steeply dipping and constitute a northeast-trending dyke swarm. The effects of episyenitization extend from several centimetres up to three metres from the dyke contacts into the medium grained monzogranite (see Table 4-2 for secondary mineralogy).

The intensity of the effects of both hematiferous and manganiferous alteration provide strong evidence in support of the view that highly oxidizing fluids were responsible for uranium mineralization. Such fluids presumably could leach uranium from the surrounding rocks and transport it as highly soluble uranyl ion. Although the source of uranium and the nature of the mineralizing fluids can only be speculated upon, meteoric fluids could have played an important part. On the other hand, the factors leading to the concentration of uranium appear to be better understood. While uranium has obviously been introduced, there is no evidence to suggest that phosphorus has also been brought in by fluids. The breakdown of K-feldspar (av. 0.21 wt.% P_2O_5) to secondary albite (av. 0.04 wt.% P_2O_5) is associated with the release of significant amounts of phosphorus to the episyenitizing fluid. Being a relatively immobile constituent, phosphorus was subsequently redeposited as secondary apatite and uranium phosphates. In this account of mineralization, uranium precipitates directly from the fluid. A change in the oxidation state of uranium from highly soluble U^{6+} -ion to the relatively insoluble U^{4+} -ion is not considered to have played a role in the concentration of uranium. It is noteworthy that pitchblende (uraninite) was not identified in this occurrence.

MANGANESE

The New Ross manganese deposits (e.g. Cain and Riddle Mines, and Marpic Shaft) are located about 12 km north of New Ross. Manganese mineralization is confined to a steeply dipping northeast-trending vein system that cuts the biotite granodiorite (Bishop and Wright, 1974). Mineralization consisting dominantly of massive pyrolusite, manganite and psilomelane occur as small veins and lenses of variable widths up to 180 centimetres (O'Reilly et al., 1982). The nature of alteration, which includes hematization and argillic alteration, is poorly understood. Weathering of the veins has produced an upper

gossan zone consisting of hematite, limonite and goethite.

Manganese mineralization has been variously interpreted as being either of supergene or hypogene in origin (Douglas and Campbell, 1942b; Bishop and Wright, 1974). In support of a hypogene origin, Bishop and Wright point to the generally high Sn and Be contents of the ore, and the presence of "hypogene-type" minerals, i.e. knebelite (iron manganese silicate) and pinakiole (oxide of manganese containing 8). These investigators suggest that manganese mineralization is genetically related to hydrothermal fluids derived during the late stages of the NRVC.

Regardless of the origin of fluids (i.e. magmatic or meteoric, or both), the possibility that the manganese was derived from the host granodioritic needs to be examined. The formation of manganese-oxide and hydroxide veins associated with the Vaughan episyenites may have been derived from the breakdown of manganiferous biotite. Thus, it is suggested that the alteration of the granodiorite along the flanks of the fracture zone could have served as a source of manganese. This hypothesis is supported by the appreciable concentration of manganese in biotites from granodioritic host rocks (range: 0.35 to 0.65 wt.% MnO) reported by Stallard (1975). In the context of a leaching model, the granodiorites with up to 20% modal biotite (McKenzie, 1974) are probably the most "fertile" of the granitoid phases in the SMB for hosting manganese mineralization.

Alternatively, manganese could have been derived by leaching of the Meguma country rocks. In view of the manganiferous contents of ilmenites reported from the Meguma Group rocks (e.g. up to 9 wt.% MnO, Cullen, 1983), the breakdown of ilmenite during fluid-rock interaction could release substantial amounts of manganese to the fluid.

OTHER TYPES OF MINERALIZATION

Although it is not the intention of this paper to summarize all the different types of mineralization recognized in the Batholith, two interesting occurrences deserve further comment.

Fluorite + Sulphide Mineralization

Fluorite mineralization has been described by Felderhof (1978) from a number of occurrences in the NRVC and adjoining region (e.g. Canoe Lake, Leminster, Walker Moly, Upper Tantalum and Reeves Tin occurrences; fluorite locality names are same as those used by Felderhof; Fig. 4-1 shows locations). He recognized that fluorite mineralization is typically structurally controlled occurring as vein infill along fractures and joints. The veins may range in width up to five cm, but are generally less than one mm in width, and discontinuous. The effects of brecciation, in which fragments of host granitoid rock are cemented with fluorite and chalcedony, have been noted by O'Reilly et al. (1982) from the Leminster occurrence. In addition to the introduction of secondary fluorite, the effects of

other types of alteration may be recognized, such as hematization, chloritization, silicification and albitization.

Partial analyses of fluoritized granitoids by Felderhof (1978) indicated strong enrichments in Cu, Pb and Zn (often greater than 100 ppm in these elements, excluding a fluoritized greisen sample). In this regard a fluoritized sample from the Upper Tantalum occurrence (sample 121; Logothetis, 1984) had a concentration of 4022 ppm Zn. Sphalerite mineralization up to several modal per cent can be recognized in some samples. In this locality, fragments of the brecciated host monzogranite are cemented together by sphalerite and purple fluorite. The extent of mineralization is not known, but it appears to be preferentially confined to zones of shearing/brecciation. In view of the widespread occurrence of fluorite-bearing granitoid rocks, fluorite + sulphide mineralization could be a common, though not previously recognized, mineralization in the NRVC.

Dunbrack Deposit

The Dunbrack Mine, a Pb-Zn-Ag deposit, is located about five km north of Musquodoboit Harbour within the Musquodoboit Batholith. MacMichael (1975) who investigated this deposit noted that mineralization is confined to a quartz fissure vein (about one metre in width) immediately adjacent the contact between a fine grained dyke body and the surrounding biotite-muscovite monzogranite. The effects of one or more episodes of silicification, brecciation, mineralization and supergene enrichment have been recognized. Evidence of chloritization (of biotite) and sericitization (of plagioclase) are evident up to about 1.5 km from the deposit. $^{40}\text{Ar}/^{39}\text{Ar}$ dating by MacMichael indicated that a time interval of about 70 Ma separated the episodes of alteration/mineralization (299.2 ± 12.5 Ma) from the crystallization of the monzogranite (368.7 ± 32 Ma) and suggested that mineralization at the Dunbrack Mine may be related to the Maritime disturbance.

CHARACTERISTICS OF GRANITOID ROCKS ASSOCIATED WITH TIN MINERALIZATION: A COMPARISON WITH THE NEW ROSS-VAUGHAN COMPLEX

In recent years there has been a growing body of literature attempting to ascertain the geological features of granitoid rocks associated with tin deposits. In general, these investigations attempt to provide one or more useful criteria for distinguishing between granitoid rocks which are not associated with mineralization (barren) from those that are (mineralized). This section considers a number of these criteria, as well as proposing a few of its own, with particular reference to the SMB. In general, the criteria are discussed in the order of decreasing generality analogous to their presumed (hypothetical) order of application, from most general criteria used in global and regional comparisons to more specific criteria that may be used to assess a granitoid body of an individual occurrence. It should be pointed out that while the discovery of the East Kemptville and Millet Brook deposits confirms the potential of the SMB with respect to

economic Sn-U mineralization, there are large regions of the Batholith which remain unmapped and whose potential is relatively unknown.

GENERAL FEATURES OF MINERALIZED GRANITOIDS

The close spatial association between tin deposits and granitoid complexes ranging in age from Precambrian to Tertiary is recognized in many tin districts of the world (e.g. Shcherba, 1970; Taylor, 1979b; Stemprok, 1979). Although the genetic relationship between tin mineralization and granitoid intrusives is not universally established, it is often assumed. Chappell and White (1974) have proposed a set of mineralogical and geochemical criteria (Table 4-3) for distinguishing between I-type granites (supposedly derived from igneous source rocks) and S-type granites (supposedly derived from metasedimentary source rocks), the latter type being associated with tin mineralization.

As indicated by Table 4-3 the SMB satisfies a number of criteria exhibited by I and S-type granites. Although the SMB, in common with other granitoid complexes associated with tin mineralization (e.g. Cornwall, Erzgebirge, Herberston and New England), is peraluminous in composition, its Na₂O values are considerably higher than that of S-type granites. In addition, the application of initial Sr isotope ratios for discriminating between I and S-type granites can give equivocal results (Flood and Shaw, 1977). In general, it appears that the only conclusion that emerges from the use of the above classification scheme is that granitoid rocks associated with tin mineralization are peraluminous in composition. However, as Clarke (1981) has pointed out, peraluminous granites may be produced by processes other than simply anatexis of metasedimentary source material. Therefore, in the light of the above comments, the application of the I and S-type classification scheme may not be suitably applied to the SMB.

ILMENITE AND MAGNETITE-SERIES GRANITOIDS

The relationship between the presence of opaque oxides in the host granitoid and its potential for tin mineralization was first pointed out by Aranykanon (1961; cited in Ishihara, 1981). He noted that granitoids hosting tin deposits contained significantly lesser amounts of magnetite compared to barren granitoids. This suggestion has been greatly expanded upon, and applied to a number of granitoid terrains in Japan, Thailand and Malaysia by Ishihara (1977), and Ishihara et al. (1979). Ishihara proposed that granitoid rocks can be subdivided into two distinct series having different mineralization potentials, i.e. "ilmenite-series" (associated with Sn-W deposits) and "magnetite-series" (associated with Cu-Pb-Zn-Mo deposits).

A compilation of the mineralogical and geochemical criteria that have been suggested as being diagnostic of the two granitoid series are listed in Table 4-4. It is evident that the granitoid rocks constituting the SMB satisfy a number of the criteria of ilmenite-series granites. As shown in Figure 4-2, the granitoid rocks of the SMB, including those associated with

polymetallic tin mineralization plot in the ilmenite-series field. However, the application of Ishihara's classification scheme to the SMB is severely limited, as the more evolved phases of the SMB do not contain either magnetite or ilmenite as a primary phase. Instead, "rutile" (the high temperature polymorph of TiO₂) is reported to be the dominant opaque oxide in these rocks, in particular of leucomonzogranites hosting mineralization potential of granitoids (Logothetis, 1981). As shown in Figure 4-2, most of the granitoids associated with polymetallic tin mineralization in the NRVC plot within the fields of granitoid rocks in which rutile is the dominant opaque. Perhaps the delineation of a "rutile-series" field may improve the general utility of the diagram in assessing the tin mineralization potential of granitoid rocks.

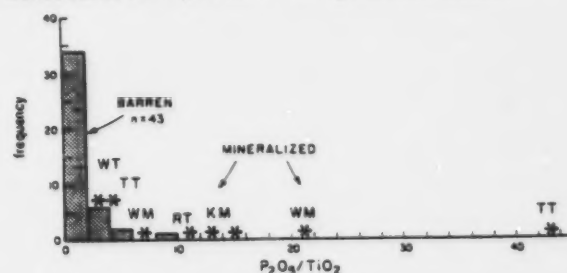


Figure 4-2. Histogram plot of barren and mineralized granitoid rocks (all lithologies) on the basis of their P₂O₅/TiO₂ ratios. Note that the barren granitoid rocks cluster in the low (0 to 4) range, whereas much higher values (up to 43) are commonly associated with the mineralized granitoids. The two highest values are associated with minor dyke bodies (i.e. pegmatite: 21 and elvan (quartz porphyry) vein: 43). With the exception of the Turner Tin prospect, all remaining host granitoids are leucomonzogranites. (Explanation: WM, Walker Moly deposit; TT, Turner Tin prospect; RT, Reeves Tin prospect; WT, Walla-back Tin prospect; KM, Kaddy Molybdenum prospect).

MAJOR ELEMENTS

In general, the search for universal diagnostic major element criteria has not been very successful for distinguishing between barren and mineralized granitoids. In a comparative study of the average compositions of tin-mineralized and tin-barren granitoids from many districts of the world, Stemprok and Skvor (1974) concluded that the differences were very small, and for most of these the observed differences were indistinguishable from that of the regional compositional variation. They suggested that the higher contents of K₂O and SiO₂ associated with tin-mineralized granitoids probably reflect the large number of analyses of leucocratic granites associated with the deposits (i.e. a sampling

Table 4-3. Geological criteria of Chappell and White (1974) applied to the granitoid rocks of the South Mountain Batholith

I-types	S-types	SMB
(1) relatively high sodium, Na ₂ O normally >3.2% in felsic varieties, decreasing to >2.2% in more mafic types	relatively low sodium, Na ₂ O normally <3.2% in rocks with approximately 5% K ₂ O decreasing to <2.2% in rocks with approximately 2% K ₂ O	Na ₂ O (95% C.I.): 3.50 - 3.80% K ₂ O (95% C.I.): 4.26 - 4.70%
(2) mol.Al ₂ O ₃ /(Na ₂ O + K ₂ O + CaO) <1.1	mol.Al ₂ O ₃ /(Na ₂ O + K ₂ O + CaO) >1.1	range (1.39 to 1.78)
(3) C.I.P.W. normative diopside or <1% normative corundum	>1% C.I.P.W. normative corundum	generally greater than 1% corundum
(4) broad spectrum of compositions from felsic to mafic	relatively restricted in composition to high SiO ₂ types	granodiorites to high siliceous leucomonzogranites
(5) regular inter-element variations within plutons; linear or near-linear variation diagrams	variation diagrams more irregular	irregular and large trace element dispersions
(6) hornblende common	biotite and muscovite common garnet and cordierite common	biotite and muscovite are common aluminosilicates present: garnet, cordierite, andalusite
(7) sphene is a common accessory	monazite	ilmenite and rutile
(8) apatite commonly as inclusions in biotite	commonly as larger discrete crystals	as inclusions in biotite and as discrete crystals
(9) mafic hornblende-bearing xenoliths	metasedimentary xenoliths common	common metasedimentary xenoliths
(10) Sr ⁸⁷ /Sr ⁸⁶ 0.704-0.706; isochrons give a regular linear set of points	Sr ⁸⁷ /Sr ⁸⁶ >0.708; isochrons show a scatter of points	range from 0.708 to 0.710
(11) associated with porphyry-type Cu and Mo deposits	associated with Sn mineralization	host to economic deposits of Sn, U and Mn

Table 4-4. The geological criteria of Mt-series and Ilm-series granites as applied to the granitoids of the South Mountain Batholith.

Granitoid Series		
Magnetite-series	Ilmenite-series	South Mountain Batholith
<u>Mineralogical characteristics</u>		
- contains magnetite: opaque oxide content >0.1 vol.% (high magnetic susceptibility)	- contains minor ilmenite mainly in ferromagnesians; opaque oxide content <0.1 vol.% (low magnetic susceptibility)	- contains minor ilmenite ± rutile mainly associated with ferromagnesians; <0.1% abundance.
commonly contains sphene and epidote of primary or secondary origin	- garnet and monazite may be present	- garnet may be present
Fe^{+3}/Fe^{+2} ratios in biotite are relatively higher in granitoids of the Mt-series than in the Ilm-series. The two series plot on either side of the Ni-NiO buffer in the $Fe^{+3}-Fe^{+2}-Mg^{+2}$ diagram		$\frac{Fe^{+3}}{Fe^{+2}} = (0.00 \text{ to } 0.61)$; Bio. plot on either side of Ni-NiO buffer
<u>Chemical characteristics</u>		
- high Fe_2O_3/FeO ratios (sub-division at approx. 0.5)	- low Fe_2O_3/FeO ratios (for granodiorite composition)	- Low Fe_2O_3/FeO ratios 90% C.I. (0.03 to 0.23)
- F, Rb, Li, Pb, Sn, and Be show little variation with differentiation	- increase with differentiation	- increase with differentiation
- positive $\delta^{34}S$ (CDT) values	- dominated by negative values	- positive values (+1 to +14‰/‰)
- low $\delta^{18}O$ values	- high $\delta^{18}O$ values	- high values (10.1 - 12 ‰/‰)
- related to Cu, Pb, Zn, Mo mineralization	- related to cassiterite and wolframite mineralization	- associated with Sn-W-Mo-Cu-Zn-U mineralization

bias), as well as the effects of silicification and K-metasomatism. On the other hand, detailed geochemical studies have been more successful in distinguishing between tin-mineralized and barren granitoids from individual tin districts (e.g. New South Wales, Juniper and Kleeman, 1979; Blue Tier Batholith, Tasmania, Groves, 1978; and Erzgebirge, Tschendorf et al., 1978). In general, these investigators note that tin mineralization is associated with highly fractionated and often potassic leucocratic phases. Flinter et al. (1972) have employed the Thornton-Tuttle Differentiation Index to distinguish between mineralization and barren granites. They noticed that rocks with D.I. > 85 are more commonly associated with tin mineralization.

In the SWB the close association between polymetallic tin mineralization and the more evolved granitoid phases is well established. On the basis of extensive field work in the NRVC, Charet (1976) found tin mineralization to be associated with the more evolved, leucocratic monzogranite bodies. In addition, Smith and Turek (1976) have used various indices of the degree of differentiation to argue that the muscovite-biotite granites and leucogranites of the NRVC and contiguous Halifax Pluton are favourable hosts for tin mineralization.

Table 4-5 compares the average composition of granitoid rocks not associated with known mineralization (i.e. "barren", n=61) with those associated with mineralization (i.e. mineralized, n=18). All mineralized granitoid rocks are from localities within the NRVC (see legend of Table 4-5 for explanations). Since leucomonzogranites constitute a large proportion of the number of granitoids associated with mineralization, the average composition of barren (n=15) and mineralized leucomonzogranites (n=11) is also shown in the above mentioned table.

Comparing barren and mineralized granitoids (all lithologies considered, columns 1 and 2) suggests a number of important major element differences. The latter are characterized by relatively lower average TiO_2 , FeO, MgO, CaO, and higher Al_2O_3 , Na₂O and P_2O_5 contents (Fig. 4-3). Computation of Student's t statistic and associated 2-tail probabilities indicates that these differences are highly significant at a probability level of 1% for TiO_2 , FeO and MgO, and very highly significant at a probability level of 0.1% for Al_2O_3 and P_2O_5 . Despite the significant major element differences recognized between mineralized and barren granitoids (all lithologies), it is unlikely that these could aid in distinguishing between these two types of granitoids. In fact the above major element differences are considered to reflect simply the fact that mineralization in the NRVC is associated with the more differentiated leucomonzogranitic bodies.

A more promising approach is to compare mineralized leucomonzogranites with barren leucomonzogranites as this rock type is the most important host of polymetallic tin mineralization. An important advantage of such an approach is the removal of sampling bias arising from the unequal representation of diverse rock types.

From this comparison a number of significant differences are apparent. The mineralized leucomonzogranites have lower average contents of TiO_2 , FeO and K₂O, and higher contents of Al_2O_3 , Na₂O, CaO and P_2O_5 than barren leucomonzogranites. However, application of Student's t-test indicates that only a few of these major element differences were significant. The differences with respect to the means of FeO and K₂O were found to be highly significant, whereas the difference in Al_2O_3 , CaO and P_2O_5 are probably significant (probability levels in the range of 5 to 10%). It is worth noting that of the two comparisons attempted above, the former (i.e. comparing all lithologies, columns 1 and 2) revealed a greater number of compositional differences between mineralized and barren granitoids. However, in the second comparison (barren vs. mineralized leucomonzogranites) a number of these differences were shown not to be as significant. This is an important conclusion as it is through comparisons of the second type (i.e. comparison of similar rock types) that major element criteria could best be employed for assessing the mineralization potential of granitoid rocks.

In view of the limited number of samples on which the above conclusions are based, a thorough discussion of these differences cannot be attempted here. These differences may be related to the interplay of magmatic and late-magmatic processes as discussed below:

Modification by Late Stage Mineralization-Alteration Processes

Although modification of major element chemistry by secondary processes may have played a part, it is unlikely that these processes have produced the major element differences observed between mineralized and barren leucomonzogranites. If the major element composition of the mineralized leucomonzogranites had been modified during greisenization, we would expect such rocks to have relatively low Na₂O contents. However, the Na₂O content of mineralized leucomonzogranites is in fact slightly higher than those that are barren.

Degree of Differentiation

This view is consistent with the relatively lower TiO_2 , FeO, MnO, MgO and CaO values associated with mineralized leucomonzogranites. Furthermore, it is relevant at this point to enquire whether the mineralized leucomonzogranites constitute a distinct type of leucomonzogranite, i.e. possessing mineralogical and textural features (e.g. grain size, reversals in the order of crystallization of the primary minerals, modal differences, presence of topaz, tourmaline, etc.) that are very different from barren leucomonzogranites. Again, more data than are presently available will be required to address such a question. In this regard, it is interesting that Burnol (1978) has pointed out that mineralization (Sn-W-Li-Nb-Ta-Be) in the French Central Massif is associated with the latest generation of leucogranites (i.e. "xenoleucogranites") which show sodium enrichment, and have mineralogical and petrographic features different from other leucogranites.

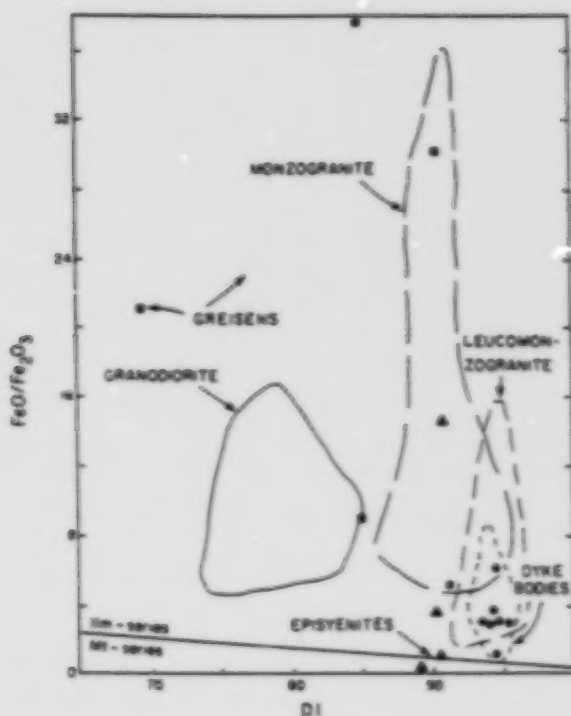


Figure 4-3. Plot of the various granitoid phases of the SMB in the $\text{FeO}/\text{Fe}_2\text{O}_3$ (rock) vs. D.I. diagram of Ishihara et al. (1979) used for distinguishing between Mv-series and Ile-series granitoids. Also plotted are samples of greisens and episyenites from thesis sample areas. Note that all samples with the exception of one episyenite fall within the ilmenite-series field. (Solid circles are mineralized granitoids).

TRACE ELEMENT CRITERIA

Trace element abundances have been extensively used as criteria for assessing the mineralization potential of granitoid bodies (e.g. Boissau-Vineau and Roger, 1980; Flinter, 1971). Underlying their application is the assumption that granitoid rocks which are hosts to economic deposits should have above average contents of Sn and other elements (e.g. F, Cl, B, Li, Rb, Cs, etc.), i.e. be specialized. Barsukov (1957), Ivanova (1963), and Juniper and Kleeman (1979) have observed that tin contents of mineralized granites are significantly higher than barren granites. Anomalous tin contents, however, are not diagnostic as some tin-mineralized granitoids (e.g. Carnarvon granite, Cornwall) are not specialized in tin (Hoeking, 1967).

Geochemical studies of the granitoid rocks constituting the SMB by Chatterjee (1980, 1981, 1982) have demonstrated that granitoid rocks hosting Sn-U mineralization are specialized with

respect to F, Sn, Li, Rb, Cs, Ta, Nb and U. Smith and Turek (1976) noted that of the three plutons they examined (i.e. New Ross, Halifax and West Dalhousie), the New Ross Pluton (i.e. NRVC) exhibited the largest enrichment (and dispersion) in F, Li, Rb and Sn, and thus was suggested to have the highest tin-mineralizing potential. Charest (1976) observed that anomalous contents of Sn, Li and Rb could be correlated with granitoid rocks associated with tin mineralization in the NRVC.

The average trace element compositions of barren and mineralized granitoids are given in Table 4-5 (columns 1 and 2). Compared with those that are putatively barren, mineralized granitoids generally have higher average contents of F, Rb, Li, Sn, Cs and Ta, and lower Ba, Sr and Th contents (Fig. 4-4). These differences are significant for Sr, Th and Ta, highly significant for Ba, and very highly significant for F, Rb, Li and Sn. The above observations support the view that tin mineralization in the NRVC is genetically related to the geochemically specialized granitoid phases, in particular specialized leucomonzogranite bodies. In general, these results are in agreement with the findings of previous investigators (Charest, 1976; Smith and Turek, 1976; Chatterjee et al., 1981, 1982; see Table 4-6). Again, it is evident that the differences between mineralized and barren leucomonzogranites are not as significant as those revealed in the above comparison (all lithologies considered). In general, the observed differences with regard to Li, Sn, Rb and Ba were found to be just significant, and for F only probably significant (at the 8% probability level).

While magmatic differentiation could explain why mineralized granitoids have higher contents of F, Li, Sn, etc., it does not adequately account for their strong enrichments and very large dispersions (Table 4-5). Although only samples exhibiting a minimum degree of alteration were used to determine the average composition of mineralized granitoids shown in Table 4-5, it is possible that their trace element compositions have been altered by fluid interaction. This interpretation is in keeping with the close spatial association of many of these mineralized granitoid samples with zones of mineralization or greisenization (or both).

Chatterjee (1981, 1982), and Chatterjee and Muecke (1982) have proposed that the granitoid rocks of the SMB can be subdivided into a "cogenetic suite" and a "paraintrusive suite" of granitoid rocks. The cogenetic suite is characterized by continuous chemical variation trends which can be attributed to the fractional crystallization of plagioclase and biotite. In contrast, the paraintrusive suite exhibits large trace element variations presumably due to the interaction of a fluid phase with residual magma or crystalline rocks (or both). Chatterjee and Muecke (1982) suggest that Sn-W, Sn-Bi, Sn-W-U, W-Mo-U and U-P-F mineralizations are associated with the paraintrusive suite. With respect to the use of trace elements as guides or indicators of mineralization potential, the results of

Table 4-5. Comparison of barren and mineralized granitoid rocks from the SMB. Columns 1 and 2 compare barren (all rock types) with mineralized granitoids (all rock types). Columns 3 and 4 compare barren and mineralized leucosomes. The (two-tailed) probability associated with the student's t is given for both comparisons. Probability levels: (*) significant ($5\% > p > 1\%$); (**) highly significant ($1\% > p > 0.1\%$); and (***) very highly significant ($0.1\% > p$).

Column	All lithologies				Leucosomes			
	1		2		3		4	
	n	barren $\bar{x}$ s.d.	Mineralized (MRVC) $\bar{x}$ n s.d.	t-test (two-tail) probability	n	barren $\bar{x}$ s.d.	Mineralized (MRVC) $\bar{x}$ n s.d.	t-test (two-tail) probability
SiO ₂	43	72.99	2.45	9 73.72 2.15	6	74.53 1.21	74.37 1.21	0.827
TiO ₂	43	0.27	0.22	9 0.04 0.04	6	0.09 0.07	0.04 0.02	0.143
Al ₂ O ₃	43	13.70	0.49	9 14.77 1.28	6	13.81 0.47	14.31 0.45	0.089
Fe ₂ O ₃	43	0.25	0.17	9 0.20 0.09	6	0.22 0.16	0.21 0.04	0.785
FeO	43	1.85	1.11	9 0.70 0.29	6	0.97 0.19	0.65 0.08	0.004**
MnO	43	0.05	0.03	9 0.04 0.02	6	0.03 0.01	0.03 0.01	1.000
MgO	43	0.43	0.39	9 0.07 0.07	6	0.14 0.11	0.06 0.02	0.126
CaO	43	0.90	0.65	9 0.53 0.17	6	0.44 0.08	0.58 0.15	0.068
Na ₂ O	43	3.56	0.46	9 4.07 0.69	6	3.86 0.67	4.36 0.46	0.168
K ₂ O	43	4.46	0.51	9 4.58 1.67	6	4.66 0.37	3.98 0.35	0.008**
P ₂ O ₅	43	0.14	0.08	9 0.33 0.08	6	0.20 0.10	0.32 0.08	0.060
H ₂ O ⁺	43	0.82	0.13	9 0.86 0.17	6	0.85 0.13	0.83 0.12	0.808
F	57	0.17	0.14	14 0.43 0.37	13	0.30 0.19	0.57 0.45	0.081
Ba ppm	60	234	250	16 62 94	14	54 102	62 112	0.035*
Sr	51	46	52	18 17 12	13	12 9	19 13	0.141
Rb	61	423	226	18 834 356	15	637 213	927 366	0.018*
Li	38	331	195	13 824 729	13	408 206	915 782	0.035*
Sn	61	25	14	13 55 20	15	308 14	54 21	0.022*
W	8	1	0.4	1 4	2	1 0	4	
Mo	36	2	0.7	13 2	13	1 1	2	
Cu	38	9	9	13 12	13	12 14	10 14	
Zn	31	60	13	0	3	55 23	0	
Th	18	9	5	5 3 1.8	3	5 1	3 1	0.247
Ce	18	23	11	5 24 3	3	41 8	24 5	0.073
Sc	18	5	4	5 2 2	3	3 0	2 2	0.394
Ta	18	4	2	5 7 4	3	8 2	5 1	0.188
Hf	18	3	2	5 2 1	3	2 0.3	2 0.03	0.974

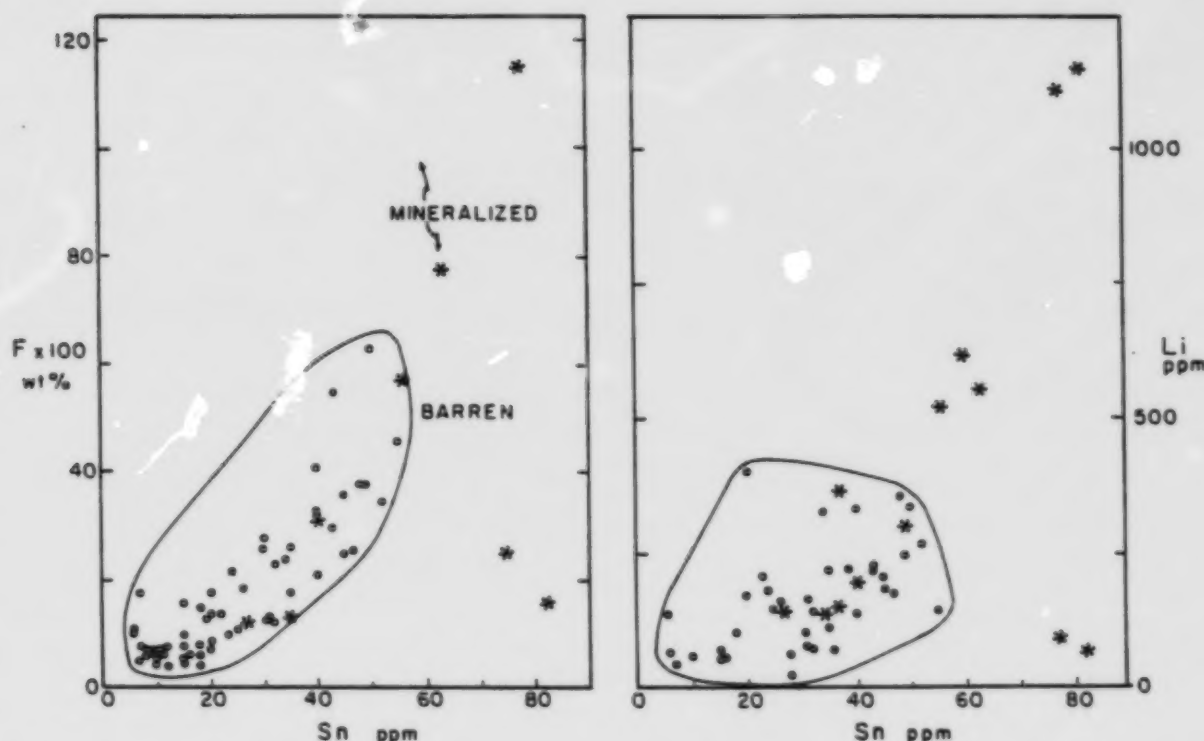


Figure 4-4. Comparison of barren and mineralized granitoid rocks plotted in the F vs. Sn and Li vs. Sn diagrams.

two distinct geochemical trends with respect to the behaviour of thorium (Th) in the SMB. A Th-depletion trend with accompanying increase in U leading to low Th/U ratios that characterizes the New Ross Complex (i.e. NRVC) and Lake George, East Dalhousie and West Dalhousie Plutons. However, in the Davis Lake and Plymouth Plutons which contain important tin mineralization there is enrichment in both thorium and uranium with increasing differentiation. In addition, they noted that areas of high thorium (≈ 6 ppm e Th) and total count in airborne gamma ray spectrometric maps could be correlated with the margins of the SMB where many of the major mineral occurrences are found. Chatterjee and Muecke (1982) proposed that granitoid bodies which show the Th-enrichment trend may be more favourable for Sn-U mineralization than areas showing Th-depletion. Additional support for this view is provided by the investigation of the greisens at Long Lake which were found to be enriched in thorium.

Chatterjee et al. (1983) have employed the statistical technique of discriminant analysis to distinguish between "stanniferous" and "uraniferous" series of specialized granitoid bodies on the basis of their major and trace element composition. Their study focused on the geochemical differences between a tin-specialized pluton (i.e. Davis Lake Pluton) which is associated with economic tin mineralization, and an uranium-specialized pluton (i.e. the East Dalhousie Pluton) and the Lewis Lake Pluton in the NRVC.

Their findings indicate that the trace elements U, Sn, Th, Li and F could discriminate between these two series of specialized granitoids.

ALTERATION AND PETROGRAPHIC CRITERIA

Assessing the mineralization potential of a granitoid body involves more than simply showing that it is specialized in certain elements. The presence of these constituents in enhanced concentrations may not be sufficient for the formation of an ore deposit. Other critical factors which are stressed as important preconditions for mineralization include the tectonic setting of the intrusion, geometry of the granite/country rock contact, nature of country rocks, development of fractures/joints in the host granitoids, etc. (e.g. Baumann, 1970; Shcherba, 1970; Yishendorf, 1977; Taylor, 1979b). Thus, lithogeochemical criteria in mineral exploration can be best applied when these and other features of the granitoid body are shown.

In this regard the alteration features of a granitoid body can provide valuable information in assessing its potential. Van de Pijpekamp (1982) has suggested that a petrographic investigation of the types and intensities of mineralogical changes exhibited by a granitoid body can be used to provide a measure of its tin potential. In general, he noted on the basis of degree and type of alteration that tin granites from the

Iberian Peninsula are more altered (i.e. possessed a higher petrographic score) than barren granites. Since Van de Pijpekamp's scheme for quantifying mineralogical changes is based on the alteration features recognized in biotite-bearing granites, such a scheme may not be well suited in the SMB where mineralization is generally found in the muscovitic leucomonzogranites. It would be interesting, however, to compare the alteration and petrographic features of barren and mineralized leucomonzogranites in the SMB.

Table 4-7 provides a list of alteration features commonly associated with altered granitoids in the NRVC. These features are based on the combined field-mineralogical-geochemical investigation of a number of alteration occurrences in the NRVC. Since the alteration features exhibited by a granitoid body can vary considerably, the list is intended to serve as a guide to better understanding the nature of alteration which is useful for assessing the potential of granitoid body. In addition, these guidelines in relating mineralogical changes which are readily observed in thin section with accompanying chemical changes may be of some value in the interpretation of geochemical data. By providing information on the nature and intensity of alteration, the systematic study of the alteration features of granitoid bodies could prove to be a valuable and an integral part of an exploration program.

SUMMARY

The principal types of mineralization recognized in the region of the New Ross-Vaughan Complex (NRVC) include:

1. polymetallic tin mineralization associated greisens,
2. molybdenum mineralization associated with pegmatitic veins,
3. uranium mineralization associated with shear/fracture zones, and
4. manganese mineralization associated with fracture zones.

In addition, two new and potentially important types of mineralization have been discovered by the author:

1. uranium mineralization (mainly torbernite and autunite) associated with lensoidal/sheet-like episyenite bodies (Vaughan occurrence),
2. fluorite + sphalerite mineralization associated with fluoritized biotite-muscovite monzogranite (Upper Tantalum occurrence).

Though no universal diagnostic criterion of tin mineralization has been found by any investigator, considerable time and effort could be saved if host granitoids which have the highest potential for tin mineralization could be distinguished from those which have lower potential. It is in this regard that the ongoing studies of the geological, mineralogical, and geochemical

features of granitoid rocks could be of valuable assistance in the search for tin deposits. The granitoid rocks constituting the NRVC, like many other granitoid complexes hosting tin deposits, satisfy many of the criteria that are considered to be favourable for mineralization:

1. presence of highly differentiated granitoid rocks, especially muscovitic leucomonzogranite bodies,
2. abundant and widespread evidence of hydrothermally altered granitoid rocks,
3. its structural setting (in part) within the roof or border (or both) zones of the Batholith,
4. the presence of geochemically specialized granitoid rocks.

Polymetallic tin mineralization in the NRVC is genetically and spatially associated with geochemically specialized leucomonzogranites. Compared with barren leucomonzogranites (and other granitoids) they are characterized by:

1. very high concentrations of Sn, F, Li, Rb and Cs, and
2. significantly lower FeO and K₂O contents.

In addition to the above list of criteria, consideration of the mineralogical, chemical and structural features of the diverse types of mineralization recognized in the NRVC (other than polymetallic tin mineralization) suggest a number of important controls favourable to mineralization:

1. structural level of intrusion-favourable sites of mineralization and alteration include roof or contact (granite/granite, granite/Meguma) (or both) zones,
2. presence of fracture or shear (or both) zones developed within the granitoid rocks or enclosing Meguma country rock (or both),
3. presence of suitable source rocks (e.g. Mn-specialized or P-specialized granitoid rocks with regard to manganese and uranium mineralization, respectively).

Table 4-7. Alteration Guides

A. Alteration features recognizable in the field

Features	Significance
(i) Greenish discoloration of granitoid rock/ - characterized by green plagioclase cores	Sericitization of plagioclase chloritization of biotite (or both) - sericite pseudomorphs after plagioclase and chlorite pseudomorphs after biotite may be common - indicative of zone of greisenization near- by (e.g. QS, QSPF, Turner Tin occurrences, and elsewhere)
(ii) Pinkish/reddish discoloration of granitoid rock/outcrop - may be related to geological contacts, joints, quartz veins, fracture zones, as well as being pervasive - characterized by "reddened" plagioc- lase, K-feldspar or biotite; fractured types may be very crumbly	Hematitic alteration - intense hematization may be characterized in thin section by one or more of the following: hematite-filled, permeating microfractures; hematite inclusions in feldspars; and hematite pseudomorphs after biotite white (or both) mica - common feature, associated with greisens, episyenites and fluoritized granitoid rock (in addition to hematized granitoid rocks)
(iii) Reddish-brown to dark brown discoloration of granitoid rock/outcrop	Manganiferous alteration - may be due to the presence of Mn-oxide/ hydroxides as inclusions or in micro- fractures in the rock - a characteristic feature associated with the Vaughan episyenites (U minerali- zation), and altered granodiorites (Mn deposits, near New Ross)
(iv) "Bleached" (white) discoloration of granitoid rock/outcrop	May be indicative of: fluoritization (e.g., Lewis Lake, Upper Tantalum); or greisenization (Long Lake); or weathering (weathered rinds on outcrop surface)
(v) Bright yellow discoloration of granitoid rock/outcrop	Supergene alteration of pyritiferous zones (e.g., QSPF greisens)
(vi) Black quartz in hand specimen samples	Chloritized quartz (i.e., Fe, Mg-metasomatism) - abundant vermiform-textured inclusions of chlorite in quartz (e.g., Turner Tin)
(vii) Vugs, open spaces and pores in granitoid rocks	May indicate intense silicification silicification (or both) (i.e. Si-metasomatism) - very common feature of greisens, episyenites, fluoritized granitoids - such porous granitoid rocks are suitable hosts for sulphide deposition in open rocks
(viii) Feldspar veins/veinlets (oriented) feldspar metacrysts (or both)	Potassic alteration - late-stage alteration-phenomena associated with some greisen zones (e.g., Long Lake, Turner Tin)
(ix) Pegmatite-looking veins (often Mn-mineral- ized)	Metasomatic "pegmatite" veins indicative of K-metasomatism - closely associated with zones of greisenization (Long Lake)
(x) Fluorite (in joints, fractures and breccia zones)	Fluoritization - indicative of fluoritized granitoids and fluoritized greisens - associated with Zn mineralization (Upper Tantalum occurrence)

Table 4-7. (Continued)

B. Petrographic guides

Petrographic Features		Significance And Associated Chemical Changes
(i) Secondary white mica	replacing primary feldspars biotite and secondary chlorite	Phengitization/muscovitization: - decrease in Na, Ca, K, Si, Ba, Sr, and REE - increase in Li, Cs, Rb, and Ta
(ii) Secondary quartz	replacing pre-existing minerals (skeletal, replacement textures common); associated with increase in modal quartz	Silicification: - strong increase in Si, accompanied by "apparent" decrease in other major elements (due to dilution)
(iii) Chlorite	(a) metasomatic-type (b) replacement-type	Metasomatic-type of chlorite indicative of Fe, Mg, Mn - metasomatism and may be associated (in addition) with increases in V, Sc and REE
(iv) Rutile	(a) metasomatic-type (b) replacement-type	Metasomatic-type of rutile indicative of Ti-metasomatism - may be associated with increases in Ti, Mn, V, Sc, Ta, Nb, REE
(v) Secondary apatite	common as abundant, minute inclusions fracture infilling (or both) within secondary albite	Indicative of albitization and apatization - at Vaughan apatization is related to the formation of U phosphates - may be associated with increase in Sr and Ce
(vi) Secondary tourmaline, fluorite, topaz		Indicative of F $\pm$ B-metasomatism fluoritization is associated with increases in F, Ca and Sr, and decrease (leaching) of metals (e.g. Li, Cu, Fe, Zn)

CHAPTER 5

GEOCHEMISTRY OF THE GRANODIORITE HOSTING
URANIUM MINERALIZATION AT MILLET BROOK

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Chatterjee, A. K., Strong, D. F., Clarke, D. B., Robertson, J., Pollock, D. and Muecke, G. K., Geochemistry of the granodiorite hosting uranium mineralization at Millet Brook; in Guide to the Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

ABSTRACT

Intrabatholithic U - Cu - P - CO₂ ± B ± F mineralization occurs in steeply dipping fracture zones in biotite granodiorite at Millet Brook. The fracture fillings contain sparsely distributed sooty pitchblende (below 50 m depth), together with pyrite, sphalerite, chalcopyrite and proustite in a gangue of quartz, calcite, phyllosilicates and tourmaline. Study of the wall rock provided detailed information on the composition of feldspars, biotite, muscovite and chlorite, and its chemical composition. The vein related alteration patterns are dominated by effects of albitization, carbonitization, silicification, K-feldspathization, hematitization and phosphatization.

The unaltered granitoid rocks are chemically homogeneous in most major and trace elements, and the chemical variations can be explained by a single magmatic process. In the altered granodiorites the variance increases, especially for SiO₂, CaO, Na₂O, P₂O₅, CO₂, U, Cu, Zn, B, F and Sn. Rare earth elements (REE) show that the unaltered rocks are characterized by negative slopes La/Yb (10 to 20) and negative Eu anomalies. The altered rocks have more gently sloping patterns, an overall drop in the LREE, an increase in the HREE and a slight decrease in the size of Eu anomalies, whereas the intermediate REE (Sm, Eu, Gd) remain more or less constant.

The U/Th ratio in the unaltered rocks ranges from 0.26 to 0.49 and the whole rock δ¹⁸⁰ ranges from 10.18 ‰ to 10.86 ‰. The U/Th ratio for the altered rocks ranges five to ten and the whole rock δ¹⁸⁰ values show an increase from 11.12 to 12.38 ‰, thus establishing a strong positive correlation between U/Th ratio and δ¹⁸⁰ values.

It is suggested that the source for the uranium in the deposit was not the host rocks and the uranium in the veins could not have been deposited by mobilizing uranium from the wall rocks. It is demonstrated that one of the late stage uraniumiferous leucogranite of the New Ross area has lost approximately 70 per cent of its original uranium content by aqueous hydrothermal and carbonate-rich fluids and that the mineralizing elements were transported as a carbonate complex. The deposition of uranium resulted from exsolution of CO₂ from the fluids and its reaction with the wall rocks.

INTRODUCTION

Since 1974, a number of small uranium prospects have been discovered in hydrothermal veins in the granitic rocks of the South Mountain Batholith (SMB). The present paper is concerned with the Millet Brook deposit, the only prospect which has been explored extensively and has striking resemblance to many European mining districts (e.g. Bernardan deposit in the LaCrouzille district of the Massif Central). The aim of the present investigation is to characterize the host rock, the alteration envelope and mineralization on the basis of detailed mineralogy and geochemistry. It also deals with the behaviour

of REE during magmatic and postmagmatic hydrothermal events, and examines the possibility of REE mobilization, uranium enrichment and interaction between the granite rocks and the mineralizing fluids of various origin on the basis of whole rock 180/160 measurements.

The Millet Brook uranium prospect is located in NTS 21A/16, approximately 50 km west-northwest of Halifax and approximately 18 km southwest of the town of Windsor (Fig. 5-1). In 1976, a helicopter-borne survey was flown by Aquitaine Company of Canada over parts of the (SMB) and a discrete anomaly was detected at Millet Brook. During 1978, detailed prospecting uncovered several new areas of uranium mineralization within the shear zones in the biotite granodiorite. In 1979, a helicopterborne radiometric survey and an airborne resistivity - EM - VLF - Mag survey was flown on a 200 metre line spacing, and ground resistivity became a routine exploration technique. During a period from 1978-1981, a total of 144 diamond-drill holes totalling 12 260 metres were drilled with total reserves estimated at about 1.0 million pounds of U₃O₈ with an average grade of 0.15 - 1.20% U₃O₈, at a cut-off grade of 0.10% over a 2.0 metre width.

ANALYTICAL METHODS AND TREATMENT OF DATA

The present study is based on a systematic study of 180 samples of which 21 were collected from the surface outcrops and the remaining from 17 diamond-drill holes from C1, C2, A9, F and BLS zones of the Millet Brook area (Fig. 5-1). The samples from the drill core (length ranged from 12-19 cm) was sliced lengthwise, one half being subsampled for thin and polished sections and then thoroughly cleaned and powdered for chemical analyzes. The remaining portion after staining with Co(NO₃)₃ was retained as a permanent collection.

Petrographic examination (texture, mineralogy and type of alteration) was made from thin and polished sections, the latter were subsequently used for microprobe and fission track studies. The composition of silicate and opaque minerals was determined by electron microprobe (Cambridge Mark V) equipped with wavelength dispersive and energy dispersive systems, and EDATA data reduction program.

The major oxide analyzes were done by using the analytical techniques of Abbey (1970), using a Jarrell-Ash 810 dual channel atomic absorption unit, except for SiO₂, which was determined colorimetrically using a lithium meta- and tetraborate fusion method.

The Rb, Ba, Sr, Be, Sn and Mo contents were determined by non-volatile and volatile methods using a 3.4 metre Ebert Jarrell-Ash grating spectrograph, and the intensities were measured on a 21300 Jarrell-Ash photodensitometer. The buffers for non-volatile and volatile methods were lithium carbonate, lithium fluoride, K₂SO₄, Eu₂O₃ and PdI₂, and Al₂O₃ + BaCO₃ and KIO₄, respectively. The limits of sensitivity are: Be, 3 ppm; Ba, 5 ppm; Sr, 3 ppm; Rb, 1 ppm; and Mo, 1 ppm. The Cu, Pb, Zn and Li concentrations were

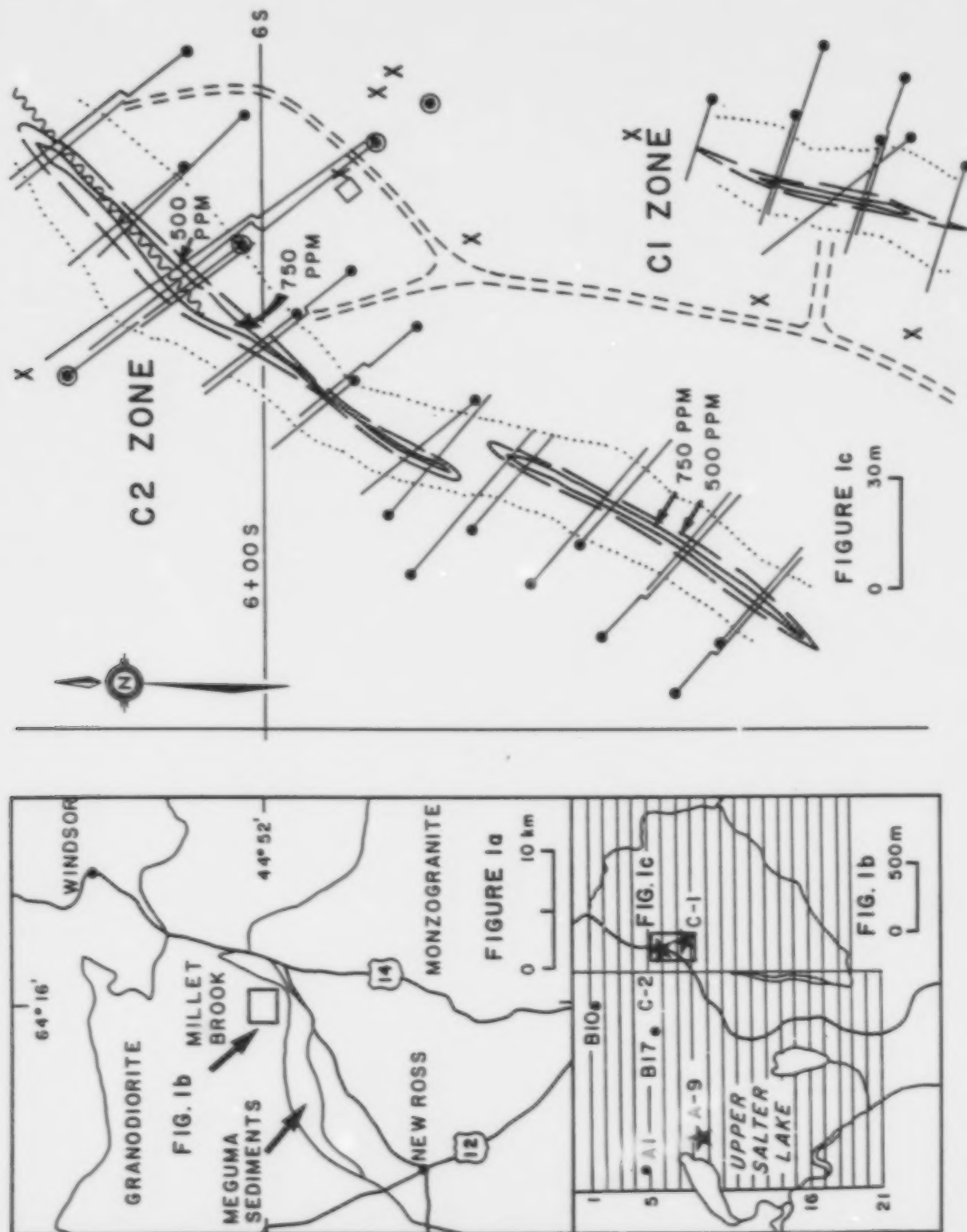


Figure 5-1. Map showing the generalized geology of the Millet Brook area. C1, C2, A9 are locations of mineralized zones in the biotite granodiorite of the South Mountain Batholith.

determined by atomic absorption, with background correction applied for Pb, with limits of sensitivity less than 2 ppm.

Fluorine content of the samples was determined using a Fisher Accumet model 425 digital pH/ion metre. Samples (0.25 g) were fused with sodium carbonate - potassium nitrate (2:1) flux and dissolved in water. The solution was mixed with 50 mL of TISAB buffer and the fluoride concentration was measured with a specific ion electrode using the method of standard addition.

The uranium and thorium contents of the rocks were determined using delayed neutron activation analysis by the Atomic Energy Commission of Canada. For uranium the samples were sealed in polyethylene capsules, stacked in the sample loader and transferred pneumatically to the reactor for 60 second irradiation in a thermal neutron flux of approximately 10^{12} neutrons $\text{cm}^{-2} \text{ s}^{-1}$. The isotopes with short half lives were allowed to decay for 20 seconds, and the samples were counted for 60 seconds. The samples were compared with standards in which the uranium concentration is known from intralaboratory determination: DHG 1-1 (4.6 ppm), DHG 1-2 (18 ppm), BL 3-6 (35 ppm), Sy-3 (876 ppm).

Thorium content of the samples was determined using neutron activation followed by gamma spectrometry. The samples were irradiated for two hours and after seven days storage the 311.9 KeV gamma peak of ^{233}Pa ($^{233}\text{Th}/^{233}\text{Pa}$) was measured using gamma spectrum measurement. Both the activation and measurement systems were calibrated using internationally recognized standards.

The precision of this major element determination is generally better than 1% except for some elements in low concentration for which a precision better than 2% was obtained. The precision of minor and trace elements at levels below 0.1% is better than 10%. The accuracy of the determination was checked by comparison with SY-2, SY-3 and G-2, and the values obtained are within 2% of the recommended values (Abbey, 1977; Flanagan, 1973).

The REE were analyzed by a thin film X-ray fluorescence technique of Fryer (1977) and by INAA technique described by Luoden (1978). The REE data are normalized to the chondrite values given by Taylor and Gorton (1977). Conventional techniques were employed for the extraction of oxygen from whole rock with bromine pentafluoride and quantitative conversion of carbon dioxide, prior to mass spectrometric analysis (Longstaffe, 1979). The isotopic data are reported as δ 180 values in per unit relative to standard mean ocean water (SMOW). Uranium distribution (fission tracks) was determined by irradiating the polished thin sections with a thermal neutron flux of 10^{11} to 10^{12} neutrons/ cm^2 and recording the induced fission tracks in a mylar (0.5 mm thick) detector which was subsequently etched with 6N NaOH solution.

The interpretation of the analytical data included various statistical methods and some of

the essential statistical terminology is introduced in this section for the benefit of those who have had no training in statistical methods. The numerical data (major and trace elements) is arranged into a frequency distribution and certain characteristics, e.g. mean ($\bar{x}$), variance (s^2), standard deviation (s), skewness (sk) and kurtosis (ku). The mean is an arithmetic mean of population (N), the variance is a measure of the manner in which the measurement fluctuates about the mean, the square root of the variance is the standard deviation, skewness is a measure of departure from a symmetrical distribution about the mean and kurtosis measures the peakedness or flatness of the frequency distribution.

The petrographic study of the granodiorite hosting the Millet Brook uranium deposit showed distinct mineralogical and chemical changes and because discriminant analysis is a well known technique (Davis and Sampson, 1966) for testing differences between populations. The statistical problem consists of developing a function based on measurements obtained on each of the samples and to test statistically whether the groups of samples are from different populations. It also helps to assign new samples to the correct population when it is not known from which of the populations the individual sample comes. Thus, the discriminant function is potentially of use for studying the hydrothermally altered rocks associated with the mineralization.

The statistic used to test whether the multivariate means of the two groups are significantly different is given by Mahalanobis generalized distance (D^2) and is a measure of the distance between two points in any number of dimensions. The SPSS program (Nie et al., 1975) allows for entry or removal of a variable in a series of steps before discriminant function is determined. When more variables are suitable for entry, a matrix of the F value is printed to test the difference between each pair of sample groups statistically.

Factor analysis attempts to evaluate the underlying structure of a data set and the analysis assumes that the original variables are linear functions of a set number of underlying controlling factors. These factors may be mineralogical or geochemical and hence useful in understanding the role of magmatic versus hydrothermal processes (cf. Clarke and Chatterjee, 1985) which probably cause the variations in a single rock unit (the granodiorite of the SMB in a relatively small area (3.5 km x 2.5 km).

The computational procedure for the R-mode factor analysis (Herman, 1960; Miesch et al., 1966; Miesch and Reed, 1979) involves four steps: (1) an $n \times n$ correlation matrix is computed; (2) the R-matrix is solved for n roots (eigenvalues) and a number of roots (P) is selected as the minimum number needed to account for a reasonable per cent of the total variance; (3) an initial matrix of factor loadings defines the principal components of the system; and (4) a final matrix of factor loading is derived from the initial matrix by the Kaiser's varimax criterion. Thus, R-mode analysis determines

variable-to-variable relations. Theoretically, factor analysis requires that the number of factors to be obtained should be known before the analysis. The use of R-mode factor analysis in the Millet Brook study corresponds with a 'fishing expedition' because the number of factors was not known before each analysis. Each factor may represent a geological process and in that case each loading in a factor represents the effect the process has on the corresponding variable. The R-mode factor analysis was carried out using SPSS program (Nie et al., 1975).

GEOLOGICAL SETTING AND NOTES ON PETROGRAPHY

The uranium mineralization occurs in the biotite granodiorite of the SMB, an upper Devonian posttectonic peraluminous granodiorite-granite complex which intrudes Cambro-Ordovician sedimentary and Siluro-Devonian volcanic rocks over an area of more than 10 000 km². During the past ten years, the SMB has been intensely studied and explored for granite hosted granophyllite mineral deposits and these resulted in more than twenty-five scientific publications and two significant deposits i.e., the tin deposit at East Kemptville (discussed in Chapters 8, 9 and 10, this volume) and the uranium deposit at Millet Brook. Clarke and Muecke (1985) have summarized various geological, mineralogical, geochemical, isotopic and metallogenetic data regarding the evolution of the SMB and the readers are referred to the above publication and the reference contained therein for detailed discussion. The exploration history, petrology and the geochemical signatures which characterize the Millet Brook deposit is available from Chatterjee et al. (1982, 1983), Robertson and Duncan (1983), and Chatterjee and Strong (1984b). The geology of the Millet Brook area is dominated by the biotite granodiorite and it is appropriate to describe the general features of this rock unit.

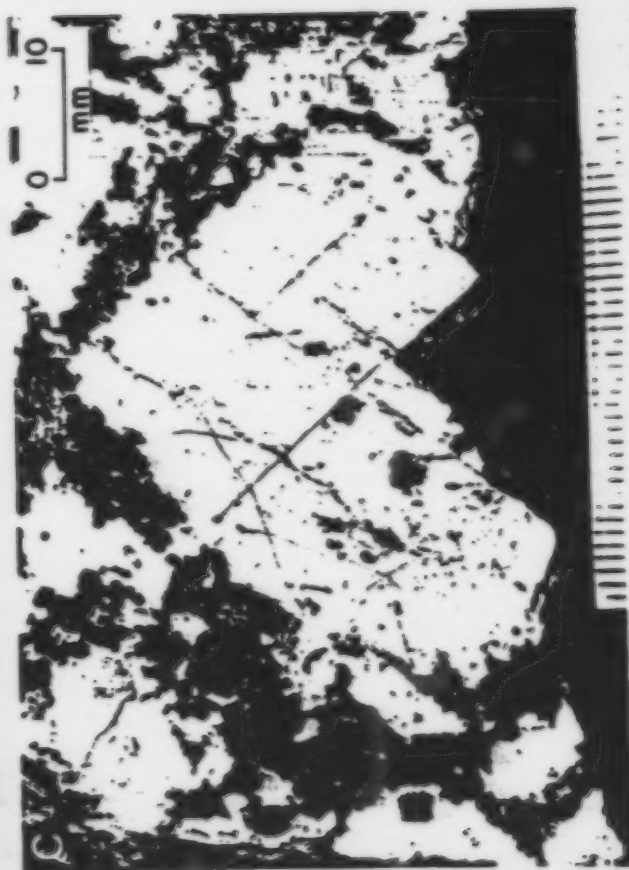
The granodiorite of the Millet Brook area is a coarse grained, coarsely porphyritic rock which constitutes part of the most volumetrically abundant rock unit of the SMB. The granodiorite is generally massive, but locally it is faintly foliated. Mafic minerals are fairly abundant and typically form aggregates that form a meshwork around the light-coloured minerals. The granodiorite, in general, has remnants of Meguma wall rock as well as scattered inclusions of uncertain origin. Near the mineralized zones, the granodiorite is intensely fractured and disaggregates readily. Phenocrysts are plagioclase and potassium feldspars, and are as much as four centimetres long. In most rock they make up about five per cent of the volume. Zones of sheared rock, and veinlets of quartz, carbonate and phosphate minerals, and quartz-carbonate-hematite-biotite-chlorite within these shear zones are also a common feature of the Millet Brook granodiorite. In the northern parts of the Millet Brook area, veinlets and dykes of leucocratic granite are common.

In this section, the granodiorite is seen to have a hypidiomorphic granular texture in which the groundmass grains are three to six millimetres in diameter. A cataclastic texture

is superimposed on the hypidiomorphic granular texture in specimens near the mineralized zones. The specimens showing mosaic or seriate texture are easily identified on the basis of bimodality of the grain size.

The main constituents of the granodiorite are quartz, two feldspars and biotite. Quartz is anhedral and often shows undulatory extinction. Plagioclase forms subhedral crystals of which approximately half have crystal cores of andesine and rims of albite, whereas the remainder are albite only. In most specimens, the plagioclase is moderately to intensely altered to sericite, calcite and clay minerals. The K-feldspar show simple Carlsbad or cross-hatched twinning but the microcline grid twinning is often blurred and discontinuous through the crystal. Most K-feldspars are perthitic with lace and patch perthite types, and coarse and fine perthite sizes equally common. Albite in perthitic grains makes up less than three per cent. Kaolinite alteration of K-feldspars is ubiquitous and, in many specimens, intense. Biotite, or its chlorite alteration mineral, is the most abundant dark mineral. It forms clusters of relatively large subequant grains or felted aggregates of relatively small tabular grains. Accessory minerals are zircon, apatite, ilmenite, garnet and the typical alteration minerals are muscovite, chlorite, tourmaline, biotite, calcite, iron oxides and smectite, and are present in various combinations.

Much of the general description and classification of the samples into various rock types of the Millet Brook area in this investigation is based upon field evidence, textural analysis and mineralogy. Special attention is paid to the grain boundary relationships between the two adjacent minerals. For example, (Fig. 3-2 to 3-4) on approaching the mineralized zone the rocks show strong effects of albitic mantling of the feldspars and hematite alteration. Often the secondary alteration is similar to surface weathering but essentially the same sequence is observed at depths. Also, the altered rocks are in sharp contact with the mineralized zone which is typically represented by uranium phosphates above the 50 metre depth. In other cases the hydrothermally altered rocks are essentially represented by a marked increase in modal muscovite and biotite + chlorite and the rocks has a vuggy, porous appearance, and where the minerals have been extensively replaced, the altered rocks are akin to pseudoporphyry. The groundmass minerals are mostly dominated by phyllosilicate minerals. As mentioned before, the mineralogy of the granodiorite is dominated by quartz, feldspars and phyllosilicates, and the accessory minerals (garnet, tourmaline, calcite, zircon, rutile, ilmenite, pyrite) generally constitute less than five per cent. Because the rock types include many varieties of hydrothermally altered rocks the minerals were classified as magmatic, relict magmatic and hydrothermal. Magmatic infers that the mineral crystallized from a magma and has undergone no appreciable alteration. Relict magmatic is used for a mineral of magmatic origin which has undergone exchange reactions with hydrothermal fluids but which closely retains its original



showing the characteristic alteration of the

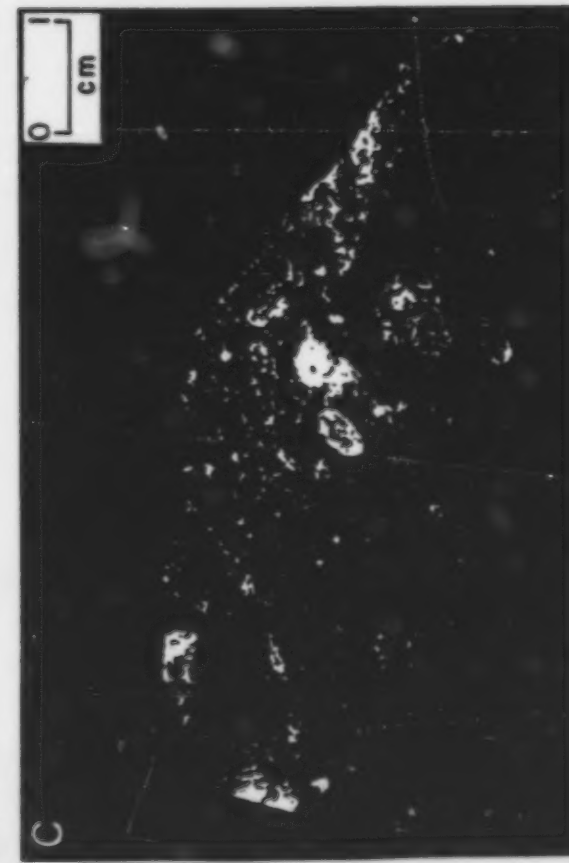
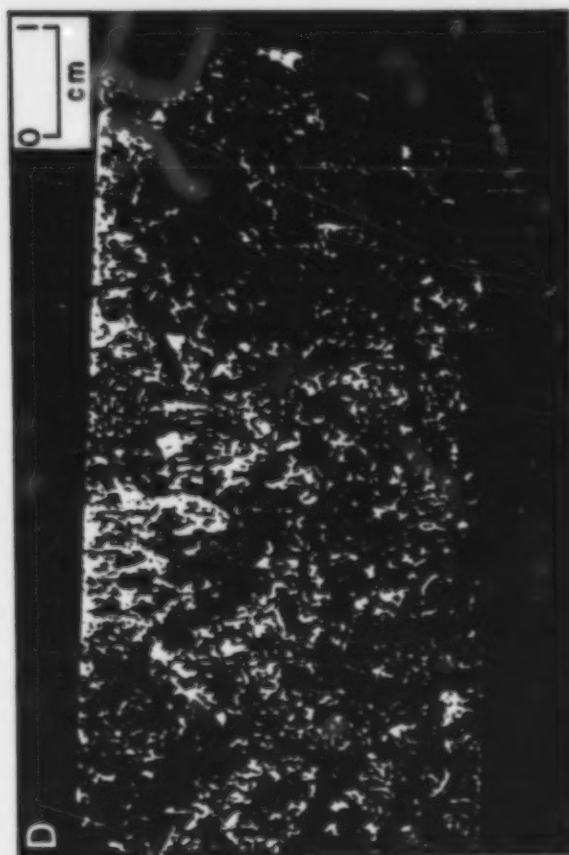


Figure 5-3. Figure showing similar relationships to that of Figure 5-2 but at a depth of 205 feet. Note the uranium mineralization along fracture in (D).

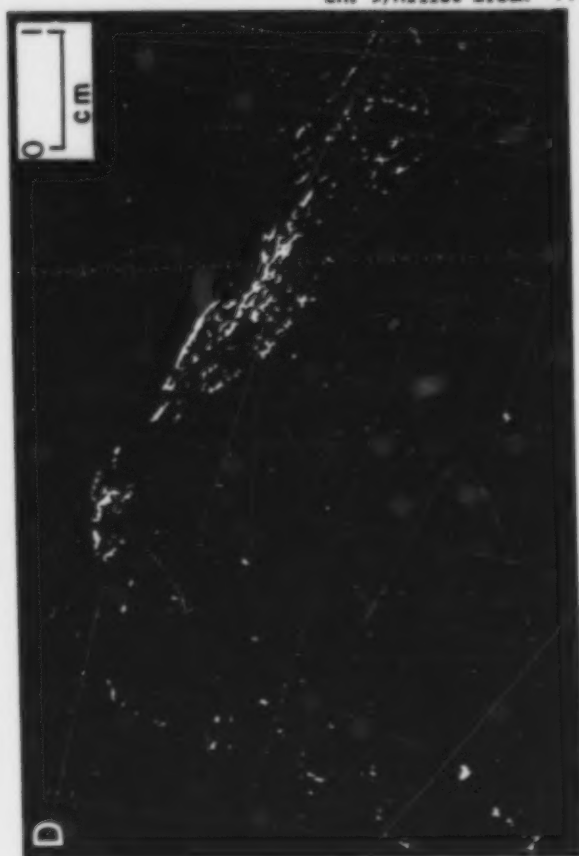
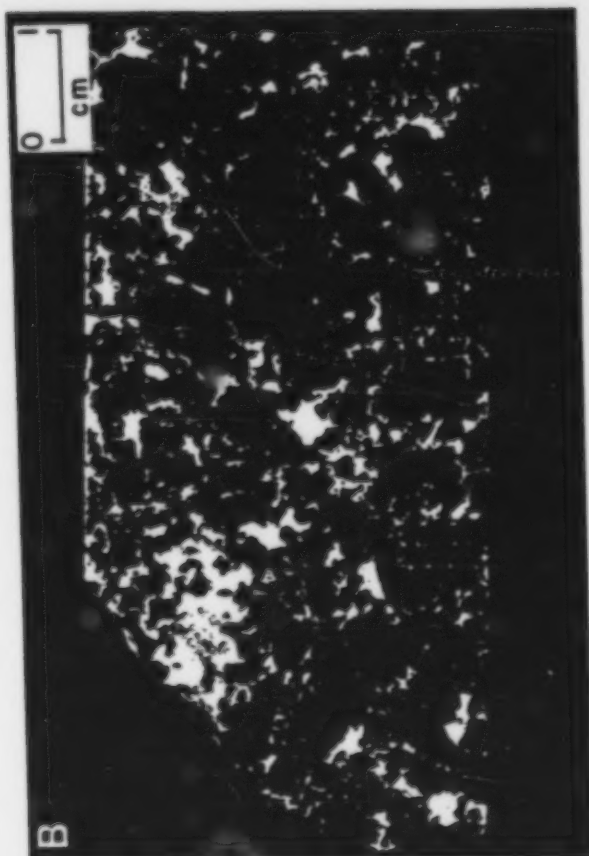


Figure 5.4. Photomicrographs of drill core illustrating the intense biotite alteration and 'porphyry' nature

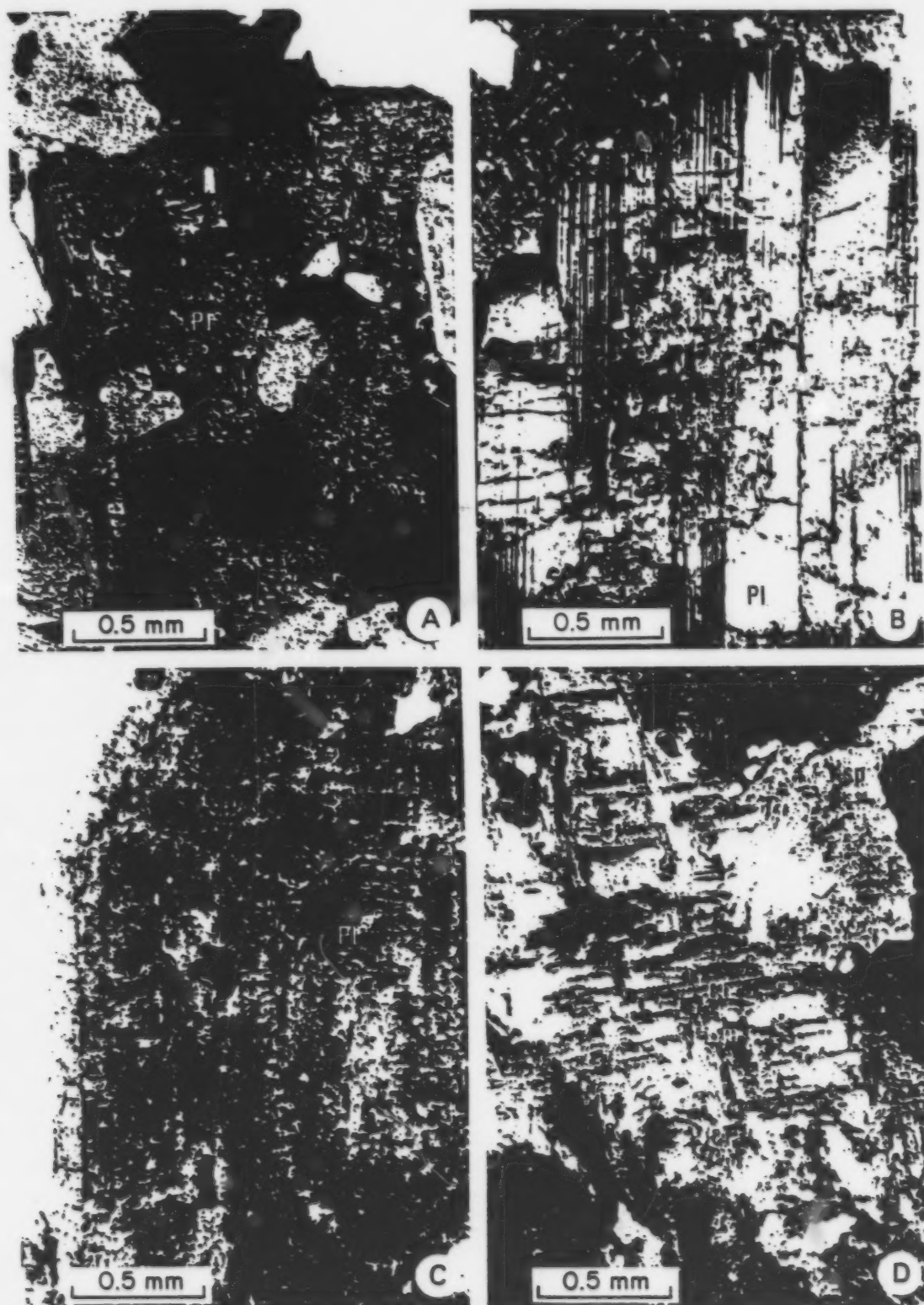


Figure 5-5. Photomicrographs showing various type of alterations of plagioclase associated with the hydrothermal activity at Millet Brook. See text for details.

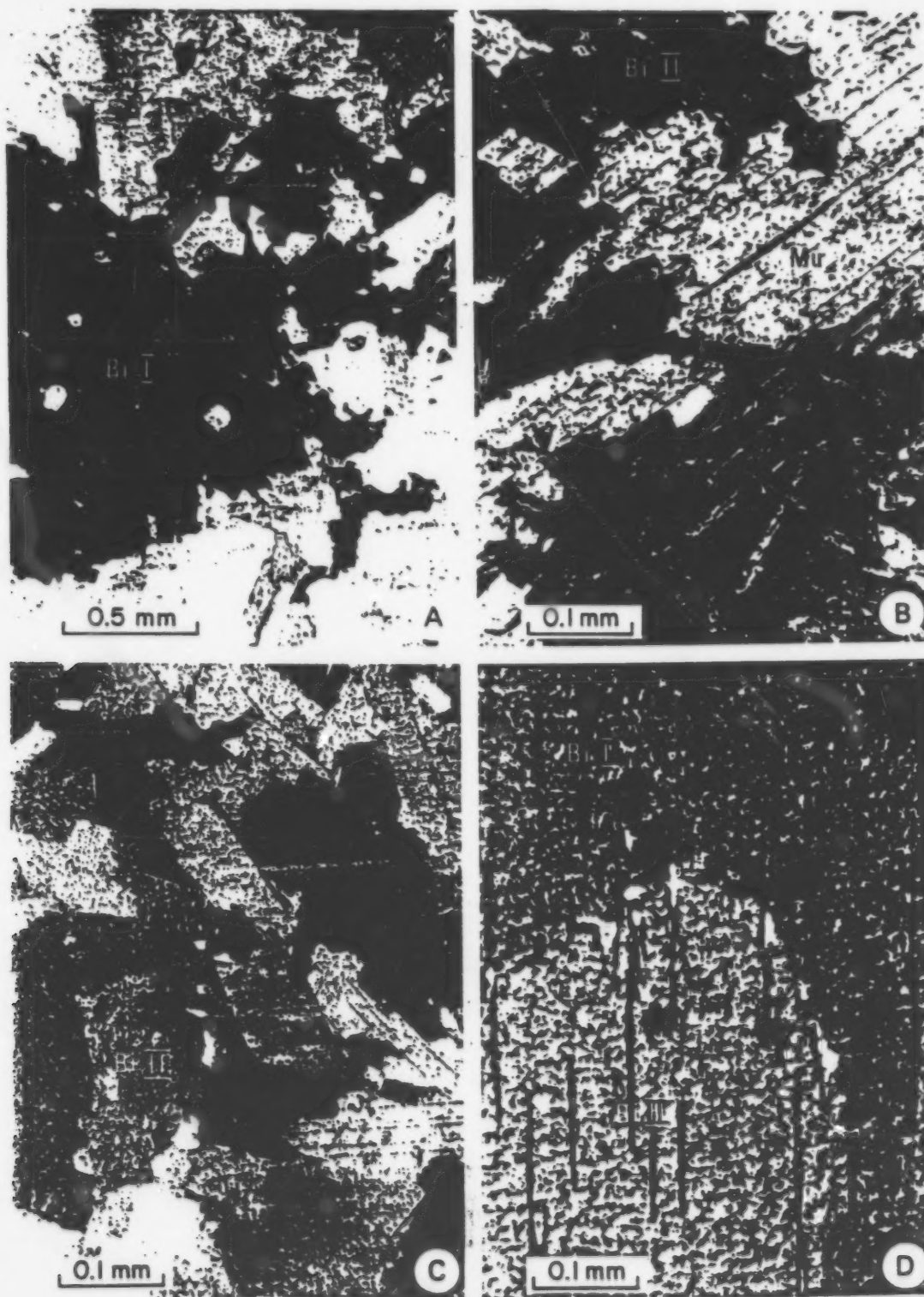


Figure 5-6. Photomicrographs showing the corrosion of earlier minerals by secondary mineral and also generation of fresh minerals during hydrothermal alteration of the granodiorite at Millet Brook.

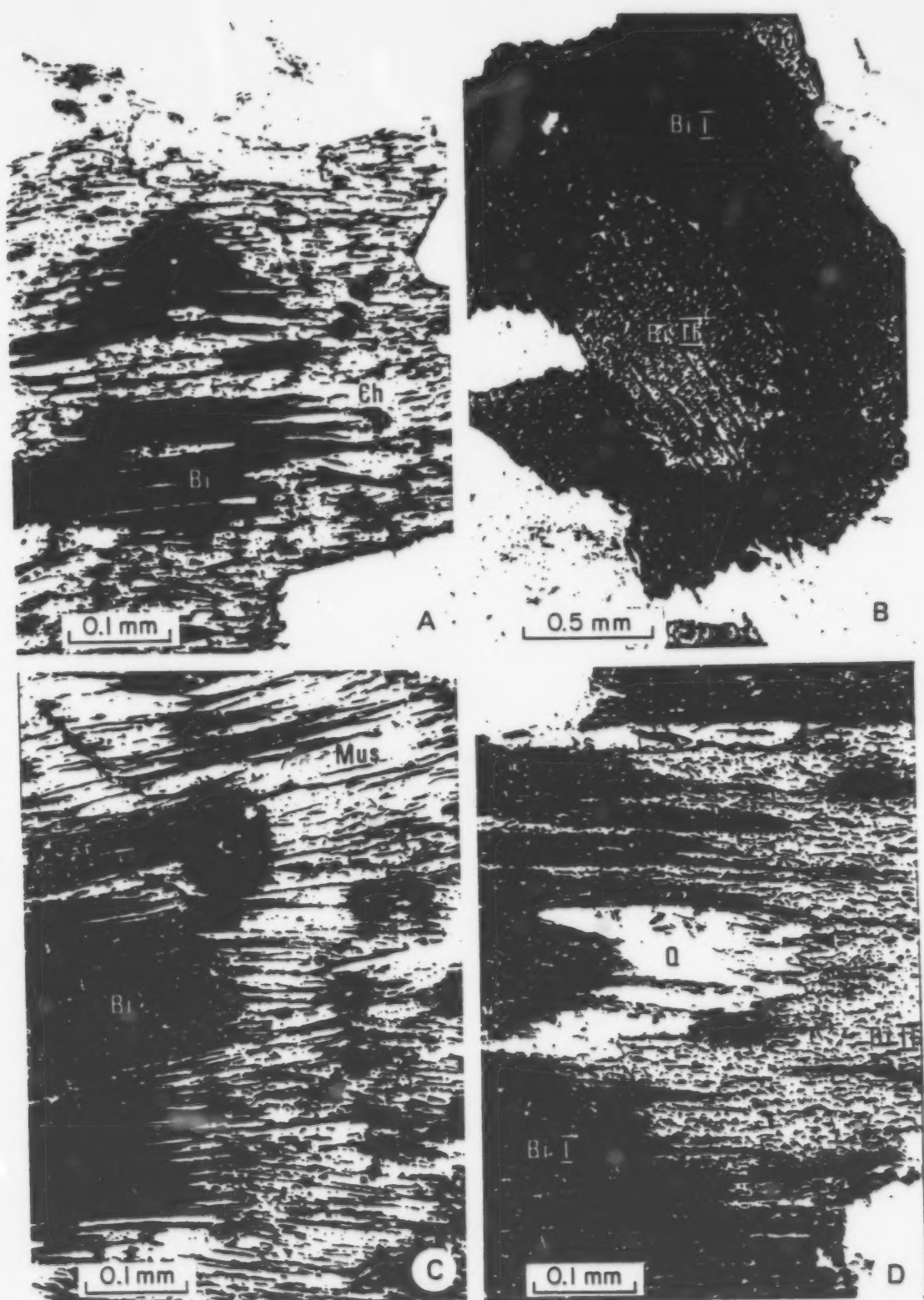
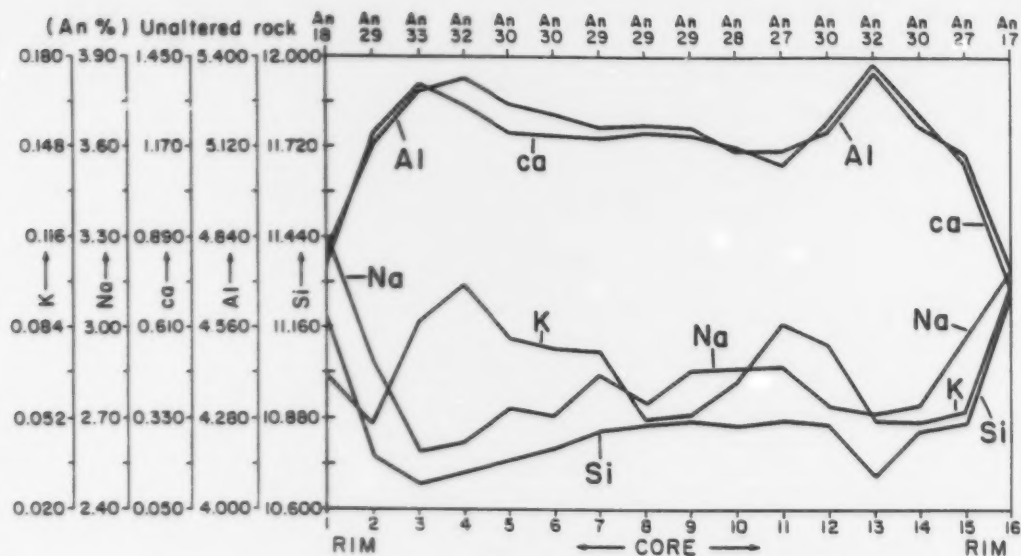
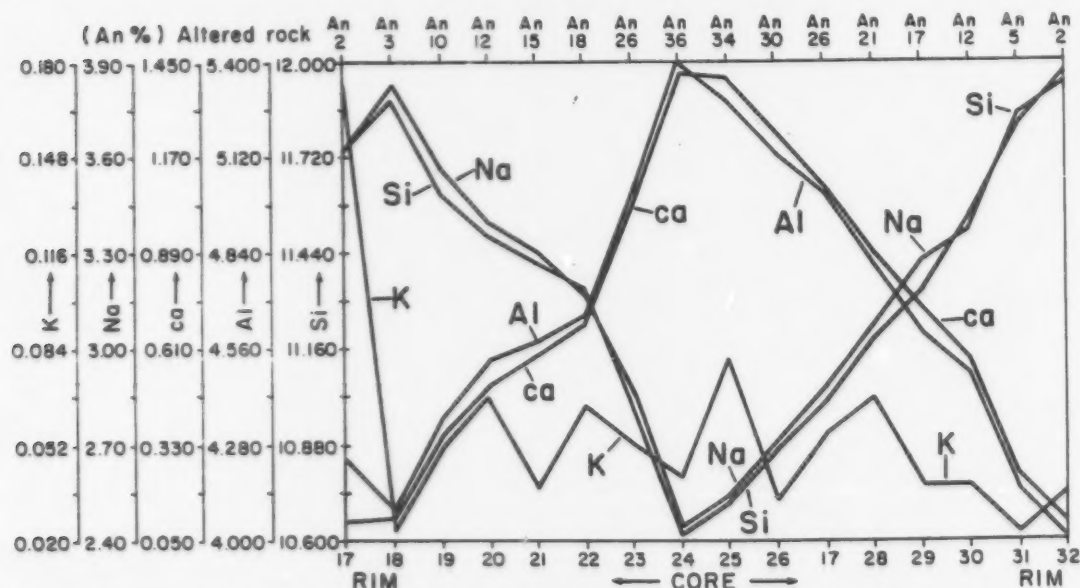


Figure 5-7. Photomicrographs showing the alteration of magmatic biotite to secondary muscovite, chlorite and also generation of secondary biotite from the primary biotite.



COMPOSITION OF THE PLAGIOCLASE FELDSPAR: UNALTERED GRANODIORITE; MILLET BK.

Figure 5-8. (A) The distribution of Si, Al, Ca, Na and K in the unit cell of a single zoned plagioclase feldspar from the unaltered granodiorite of the Millet Brook area. Appropriate An content of the plagioclase composition is given at the top horizontal axis and the number at the bottom horizontal axis correspond to the microprobe spot analysis (see Table 5-1 for analytical data).



COMPOSITION OF THE PLAGIOCLASE FELDSPAR: ALTERED GRANODIORITE; MILLET BK.

(B) The distribution of Si, Al, Ca, Na and K in the unit cell of a single zoned plagioclase feldspar from the altered granodiorite of the Millet Brook area (analytical data in Table 5-1).

No.	SiO ₂	Al ₂ O ₃	CaO	Na ₂ O	K ₂ O	E	Si	Al	Ca	Na	K	Z	X	Or	Ab	An
1	62.40	22.64	4.05	9.58	0.30	99.05	11.182	3.774	0.777	2.213	0.068	15.956	4.169	1.643	79.731	18.627
2	60.12	24.20	6.16	8.26	0.21	98.95	10.822	5.133	1.188	2.883	0.048	15.955	4.119	1.171	69.987	28.843
3	60.45	35.63	7.11	7.66	0.38	101.23	10.658	5.214	1.343	2.618	0.085	15.982	4.047	2.112	64.701	33.187
4	60.67	25.47	6.90	7.82	0.44	101.30	10.689	5.288	1.302	2.671	0.099	15.977	4.073	2.428	65.590	31.982
5	61.05	25.26	6.47	8.12	0.35	101.25	10.746	5.240	1.220	2.771	0.079	15.926	4.070	1.931	68.088	29.981
6	60.66	25.03	6.44	7.93	0.34	100.40	10.761	5.232	1.224	2.728	0.077	15.994	4.029	1.910	67.705	30.385
7	61.18	24.97	6.33	8.20	0.33	101.01	10.789	5.189	1.196	2.804	0.074	15.978	4.074	1.822	68.820	29.358
8	61.02	24.91	6.27	8.15	0.22	100.57	10.797	5.194	1.189	2.796	0.050	15.991	4.034	1.231	69.305	29.464
9	60.80	24.61	6.26	8.30	0.23	100.20	10.808	5.155	1.192	2.861	0.052	15.963	4.105	1.270	69.685	29.044
10	61.04	24.28	6.16	8.32	0.28	100.08	10.860	5.090	1.173	2.870	0.064	15.951	4.108	1.547	69.867	28.586
11	61.13	24.34	5.83	8.37	0.38	100.05	10.874	5.102	1.111	2.887	0.086	15.975	4.084	2.111	70.682	27.207
12	61.08	24.92	6.38	8.06	0.32	100.76	10.794	5.189	1.208	2.762	0.072	15.984	4.042	1.785	68.327	29.888
13	59.96	25.67	6.95	7.97	0.23	100.78	10.622	5.359	1.319	2.738	0.052	15.981	4.109	1.265	66.628	32.107
14	61.03	24.81	6.45	8.11	0.22	100.62	10.799	5.173	1.223	2.782	0.050	15.973	4.055	1.225	68.618	30.150
15	61.56	24.42	5.84	8.66	0.24	100.72	10.877	5.085	1.106	2.967	0.054	15.962	4.126	1.211	71.896	26.793
16	63.94	22.63	3.72	9.36	0.43	100.08	11.290	4.709	0.704	3.204	0.097	15.999	4.005	2.418	80.010	17.572
17	65.84	20.18	0.46	10.44	0.76	97.68	11.806	4.264	0.088	3.630	0.174	16.070	3.892	4.466	93.263	2.271
18	68.03	19.77	0.67	11.48	0.12	100.07	11.900	4.075	0.126	3.893	0.027	15.975	4.046	0.662	96.234	3.104
19	65.73	21.12	2.09	10.40	0.23	99.57	11.603	4.393	0.395	3.560	0.052	15.937	4.007	1.293	88.841	9.866
20	65.16	21.74	2.57	9.93	0.30	99.70	11.499	4.521	0.486	3.398	0.068	16.020	3.951	1.709	85.992	12.299
21	64.75	22.06	3.20	9.70	0.17	99.88	11.419	4.584	0.605	3.317	0.038	16.004	3.960	0.966	83.764	15.271
22	63.98	22.40	3.86	9.25	0.29	99.78	11.320	4.670	0.732	3.173	0.065	15.991	3.970	1.648	79.921	18.430
23	61.80	23.97	5.46	8.28	0.23	99.74	10.986	5.021	1.040	1.854	0.052	16.008	3.946	1.322	72.323	26.355
24	59.33	25.67	7.51	7.04	0.20	99.75	10.604	5.406	1.438	2.439	0.046	16.010	3.923	1.162	62.181	36.656
25	60.01	25.30	7.20	7.26	0.36	100.13	10.684	5.128	1.187	2.714	0.034	16.000	3.935	2.064	63.264	34.672
26	61.04	24.43	6.22	7.86	0.15	99.94	10.984	5.005	1.044	2.899	0.082	15.992	3.961	0.866	68.972	30.162
27	61.86	23.92	5.49	8.42	0.25	99.70	10.872	5.128	1.187	2.714	0.057	15.989	4.000	1.416	72.472	26.113
28	62.75	22.91	4.41	8.95	0.30	99.32	11.180	4.810	0.842	3.092	0.068	15.990	4.002	1.704	77.259	11.037
29	64.60	22.23	3.60	9.63	0.18	100.24	11.370	4.610	0.679	3.286	0.040	15.980	4.005	1.009	82.042	16.949
30	65.67	21.66	2.62	9.93	0.10	100.06	11.535	4.483	0.493	3.382	0.040	16.018	3.915	1.030	86.376	12.594
31	67.87	20.25	1.17	11.04	0.10	100.43	11.831	4.160	0.219	3.731	0.022	15.991	3.972	0.560	93.939	5.502
32	68.05	19.53	0.53	11.32	0.16	99.59	11.947	4.04	0.100	3.853	0.036	16.003	3.989	0.898	96.602	2.499

Table 5-1. Microprobe analyses of plagioclase feldspar from unaltered granodiorite (analyses number 1 to 16) and from altered granodiorite (analyses number 17 to 32).

[illegible]

SAMPLE NO	34.5	80-10-8	80-10-44	80-11-14	80-11-58	80-10-45	F1-1-5	F1-1-10	F1-1-15	F1-1-20	F1-1-25	F1-1-27.
B102	45.14	48.92	47.21	46.59	47.18	46.22	71.20	73.04	71.04	73.44	74.15	75.74
V102	0.82	0.72	0.59	0.71	0.81	0.75	0.40	0.22	0.47	0.30	0.24	0.10
AL2O3	16.22	15.31	15.71	15.58	15.44	15.90	14.15	13.54	14.84	12.84	13.57	12.81
Fe2O3	0.43	0.38	0.48	0.40	0.38	0.40	0.59	0.44	0.31	0.36	0.27	0.49
FeO	3.02	3.21	3.61	3.43	3.24	4.09	2.63	2.58	2.83	2.38	2.43	0.40
MnO	0.04	0.04	0.07	0.09	0.07	0.08	0.04	0.04	0.04	0.04	0.04	0.03
MgO	1.05	1.15	1.15	1.19	1.11	1.34	0.79	0.79	0.84	0.75	0.70	0.19
CaO	1.23	2.09	1.52	1.77	1.36	1.35	0.79	0.78	0.97	0.75	1.13	0.31
Na2O	3.31	3.42	3.43	3.13	3.16	3.32	3.11	3.12	3.33	3.07	3.02	2.63
K2O	5.95	4.04	4.44	5.42	5.24	4.48	5.44	4.86	4.42	3.84	4.04	6.04
M2O+	1.41	0.84	1.07	0.93	0.92	1.30	0.90	0.77	1.05	1.14	0.84	0.45
M2O-	0.11	0.13	0.03	0.13	0.13	0.17	0.03	-0.01	1.05	0.10	0.05	0.07
F2O5	0.23	0.19	0.03	0.21	0.18	0.24	0.03	0.21	0.23	0.23	0.24	0.12
CO2 OR LOI	0.05	0.12	0.09	0.47	0.04	0.07	0.05	0.04	0.02	0.02	0.03	0.03
TOTAL OXIDES	99.23	100.49	99.58	100.05	100.37	100.11	100.35	99.35	100.75	99.40	100.33	99.43
NORMATIVE MINERALOGY (PERCENT WEIGHT)												
QUARTZ	18.35	24.34	24.00	22.00	23.39	23.40	28.71	34.03	29.42	37.21	34.47	37.19
CORUNDUM	2.84	2.21	2.14	2.94	2.84	3.49	2.35	2.59	3.24	2.40	2.83	1.75
ORTHOCASE	36.05	24.24	24.63	32.35	31.24	28.07	32.34	27.44	27.45	23.16	24.04	36.01
ALBITE	28.74	29.08	29.49	24.68	26.84	28.40	24.41	24.74	26.22	24.34	25.50	22.38
ANDRTHITE	4.34	8.41	5.95	4.55	3.37	4.81	2.24	2.47	3.21	3.40	3.79	0.43
ENSTATHITE	2.64	2.55	2.89	2.97	2.74	3.38	1.94	1.49	2.21	1.90	1.22	0.48
FERROSILITE	4.18	4.51	5.44	5.00	3.59	4.00	3.78	3.79	4.28	3.72	3.74	0.05
MASHNETITE	0.94	0.52	0.71	0.59	1.72	0.90	0.84	0.34	0.45	0.53	0.39	1.00
ILMENITE	1.21	1.34	1.12	1.26	1.74	1.45	0.77	0.40	0.90	0.58	0.46	0.19
APATITE	0.55	0.44	0.42	0.47	0.42	0.57	0.47	0.05	0.54	0.54	0.33	0.24
CALCITE	0.12	0.27	0.19	1.08	0.07	0.14	0.11	0.07	0.02	0.02	0.11	0.07
TRACE ELEMENTS (PPM)												
BA	1350	700	490	900	910	470	520	330	470	390	400	140
CU	83	2	3	12	28	8	52	16	0	14	4	15
F	420	420	380	560	520	470	480	450	400	490	430	240
MO	1.7	1.7	1.5	1.6	1.5	2.4	0.8	1.2	0.8	2.8	1.0	1.0
PP	22	14	44	34	48	44	14	10	12	14	18	20
RB	173	155	184	190	154	221	188	172	162	143	148	191
SR	4.2	4.5	4.6	3.4	3.2	4.0	7.8	7.8	7.1	4.3	5.9	5.7
SR	170	135	150	162	175	125	93	73	125	77	105	52
ZH	100	78	140	129	243	410	139	91	117	120	91	59
U	5.3	4.3	9.8	3.7	12.4	19.4	3.7	5.1	7.5	4.9	5.0	21.7
W	12	4	-4	-4	-4	-4	-4	-4	-4	-4	-4	-4
W	15	4	9	30	14	9	14	15	35	53	24	9
PC	2.8	2.3	2.9	3.7	4.1	1.7	3.5	23.0	4.2	3.4	4.7	2.9
BI	6.04	0.04	0.05	0.07	0.07	0.04	-	-	-	-	-	-
LI	75	76	83	90	72	99	117	90	110	99	92	27
SB	-1	-3	-3	-3	-3	-3	-	-	-	-	-	-
CO	4	5	7	8	7	7	-	-	-	-	-	-
NI	4	7	11	12	10	15	-	-	-	-	-	-
TH	11	4	10	9	10	10	8	7	8	9	8	-2
ZR	---	310	140	550	300	370	245	240	210	340	245	135
AS	---	---	---	.10	.60	---	-.20	-.20	-.20	-.20	-.20	-.20

SAMPLE NO	F1-7-30	F1-7-35	F1-7-45, 33F1-7-55	F1-7-58, 44F1-7-60	F1-7-70	F1-7-70, 33F1-7-85	F1-7-90
BI02	49.44	71.34	49.04	72.54	48.67	73.44	48.67
CU	0.50	0.59	0.49	0.10	0.75	0.10	0.75
FE01	15.52	13.07	14.67	14.11	14.44	14.11	14.44
FE03	0.22	0.47	0.30	0.21	0.40	0.21	0.40
FE04	3.04	3.25	3.40	1.30	1.81	1.30	1.81
FE05	0.05	0.07	0.07	0.04	0.04	0.04	0.04
FE06	0.97	1.10	1.20	0.40	1.31	1.04	1.31
FE07	1.44	1.13	1.23	0.84	2.00	1.04	2.00
FE08	3.17	2.53	3.19	3.43	3.35	3.43	3.35
FE09	4.81	4.58	4.03	4.40	4.84	4.41	4.84
FE10	4.81	4.58	4.03	4.40	4.84	4.41	4.84
FE11	0.91	1.24	0.94	0.94	0.94	0.94	0.94
FE12	0.07	0.08	0.08	0.03	0.03	0.03	0.03
FE13	0.18	0.21	0.22	0.20	0.23	0.18	0.23
FE14	0.07	0.03	0.17	0.42	0.09	0.07	0.09
FE15	100.42	100.84	99.85	100.42	100.42	100.42	100.42
TOTAL DRINKS	100.42	100.84	99.85	100.42	100.42	100.42	100.42
QUARTZ	26.26	33.86	28.09	32.70	33.54	29.48	33.54
CHROMIUM	2.46	3.27	2.54	3.27	2.93	2.50	3.27
BRINOLCLARE	28.57	27.21	26.03	27.91	28.89	21.17	28.89
ALBITE	24.93	21.59	23.12	29.24	28.40	26.94	28.40
AMPHIBITE	6.67	4.19	6.98	0.83	1.19	0.02	1.19
HEMISTATE	2.49	2.74	3.21	1.33	1.20	1.67	1.20
FERROSILITE	4.64	4.70	5.29	2.24	2.15	3.04	2.15
MAGNETITE	0.34	0.69	0.54	0.19	0.29	0.88	0.29
ILMENITE	0.95	1.11	1.32	0.31	1.45	0.89	1.45
APATITE	0.40	0.47	0.49	0.47	0.63	0.42	0.63
CALCITE	0.14	0.07	0.39	1.42	0.46	0.14	0.46
TRACE ELEMENTS (PPM)							
BA	1000	840	700	660	470	520	520
CU	14	20	10	12	27	13	13
FE	480	340	520	300	330	540	330
NO	1.6	2.2	4.0	3.0	2.2	5.1	5.1
PD	50	34	50	40	26	24	24
RB	174	207	158	115	140	154	154
SM	5.3	5.4	5.5	4.1	142	4.9	142
SR	128	128	128	128	138	145	145
ZN	154	110	231	236	310	60	310
U	4.9	19.2	3.7	4.7	5.4	2.8	4.7
W	4	4	4	4	4	4	4
BE	14	14	14	47	41	20	47
DE	2.8	2.5	4.5	5.7	4.1	2.5	5.7
PI	100	100	109	30	24	56	30
LI	100	100	109	30	24	56	30
SD	100	100	109	30	24	56	30
CO	100	100	109	30	24	56	30
NI	100	100	109	30	24	56	30
MI	100	100	109	30	24	56	30
TH	100	100	109	30	24	56	30
ZK	275	290	320	315	285	140	320
TOTAL	100.42	100.84	99.85	100.42	100.42	100.42	100.42

SAMPLE NO	F1-6-25	F1-6-30	F1-6-35	F1-6-45	F1-6-50	F1-6-55	F1-6-60	F1-6-64.10F1-6-65	F1-7-10	F1-7-15	F1-7-20
8102	71.61	71.02	77.35	49.02	48.47	47.51	70.95	72.78	49.21	72.61	70.00
Y102	0.37	0.38	0.14	0.28	0.57	0.47	0.45	0.17	0.50	0.52	0.40
AL2O3	13.84	14.83	11.99	15.73	15.38	15.42	14.15	14.43	14.88	14.05	14.28
Fe2O3	0.42	0.34	0.34	0.71	0.44	0.45	0.50	0.28	0.37	0.24	0.40
FeO	2.46	2.50	1.14	2.33	3.14	3.81	3.2	1.47	3.88	3.00	3.88
MnO	0.04	0.04	0.03	0.06	0.04	0.08	0.06	0.03	0.07	0.06	0.07
MgO	0.74	0.80	0.35	0.95	1.12	1.29	1.57	0.63	1.27	0.99	1.29
CaO	0.94	1.02	0.64	1.18	1.31	1.91	1.37	0.55	1.57	1.23	1.43
Na2O	2.93	3.37	2.64	3.53	3.39	3.43	2.97	3.08	2.93	2.89	3.04
K2O	5.06	4.56	4.78	4.48	4.74	4.14	3.81	4.92	4.03	4.08	3.51
H2O+	0.86	1.04	0.54	1.22	1.02	0.69	1.19	0.89	0.74	0.69	1.27
H2O-	0.12	0.17	0.01	0.19	0.78	0.18	0.23	0.14	0.20	0.12	0.24
P2O5	0.19	0.24	0.12	0.15	0.20	0.20	0.20	0.23	0.22	0.19	0.23
CO2 OR LOI	0.09	0.05	0.09	0.05	0.10	0.34	0.07	0.03	0.02	0.02	0.02
TOTAL OXIDES	99.71	100.46	100.24	99.88	100.72	100.32	100.49	99.61	100.52	100.49	100.28

NORMATIVE MINERALOGY (PERCENT WEIGHT)

QUARTZ	31.55	29.70	41.43	26.91	25.17	24.83	33.21	34.22	30.39	34.44	32.32
CORUNDUM	2.52	3.17	1.71	3.48	3.02	3.34	3.09	3.74	4.94	3.14	3.51
ORTHOCASE	30.33	27.14	28.38	26.87	28.34	24.57	22.76	29.47	25.51	24.18	21.04
ALBITE	25.12	28.89	25.51	30.35	29.01	29.19	24.53	26.36	28.49	24.50	26.04
ANORTHITE	2.95	3.26	2.04	4.58	4.61	6.00	6.16	1.05	0.32	4.75	5.55
ENSTATITE	1.92	2.00	0.85	2.38	2.80	3.21	2.79	1.57	2.76	2.45	3.23
FERROSILLITE	3.49	3.40	1.44	3.39	4.41	5.68	5.29	2.28	4.37	4.58	5.99
MAGNETITE	0.60	0.88	0.52	1.05	0.45	0.44	0.73	0.40	1.03	0.33	0.59
ILMENITE	0.69	0.73	0.52	0.52	1.10	1.28	0.84	0.31	1.24	0.97	1.14
APATITE	0.45	0.54	0.35	0.35	0.47	0.47	0.47	0.54	0.47	0.44	0.54
CALCITE	0.18	0.11	0.12	0.12	0.23	0.70	0.14	0.07	0.11	0.02	0.02

TRACE ELEMENTS (PPM)

BA	310	330	270	340	830	450	540	580	470	510	420
CU	18	18	4	2	7	8	8	55	13	19	29
F	480	490	280	380	490	560	480	310	370	480	560
MO	1.7	1.1	2.3	1.9	4.2	3.0	2.5	1.7	1.4	2.7	2.2
PB	16	17	17	28	34	28	27	44	18	24	22
RB	172	202	173	142	175	160	157	126	156	167	190
SK	5.1	4.7	4.7	4.2	20.0	4.1	5.0	4.4	8.3	4.7	4.7
BR	88	107	108	113	119	180	141	138	110	112	138
ZN	74	137	75	82	100	203	140	103	203	116	103
U	11.4	5.6	73.9	39.0	9.6	6.1	9.8	11.4	12.8	3.9	22.4
U	-4	-4	-4	-4	-4	-4	-4	-4	-4	-4	-4
B	65	30	18	34	34	25	42	34	49	31	34
BE	4.9	4.6	4.6	6.9	5.8	3.8	2.9	3.0	2.1	3.2	2.8
BI	-	-	-	-	-	-	-	-	-	-	-
LI	70	98	51	87	104	100	79	37	50	91	130
SB	-	-	-	-	-	-	-	-	-	-	-
CO	-	-	-	-	-	-	-	-	-	-	-
NI	-	-	-	-	-	-	-	-	-	-	-
TM	8	8	4	7	10	11	12	3	10	9	12
ZR	235	280	285	185	200	240	470	225	235	225	390
AO	-20	-20	-20	-20	40	30	-20	1.60	-20	-20	-20

SAMPLE NO	F1-1-30	F1-1-40	F1-1-45	F1-1-50	F1-1-55	F1-1-60	F1-1-65	F1-6-5	F1-6-10	F1-6-15	F1-6-17, 25F1-6-20
8102	71.91	71.49	72.51	72.04	72.84	71.45	73.15	71.25	71.04	72.45	75.43
Y102	0.25	0.42	0.30	0.23	0.31	0.31	0.22	0.48	0.30	0.48	0.43
AL2O3	18.01	14.22	13.48	13.54	13.59	13.85	13.83	13.94	13.74	14.08	14.43
FE2O3	0.12	0.06	0.94	1.07	0.31	0.85	0.33	0.62	0.37	0.34	0.22
FeO	2.90	2.55	1.81	2.02	2.56	2.02	2.44	2.94	2.58	2.54	3.04
MNO	0.07	0.04	0.04	0.07	0.04	0.05	0.05	0.07	0.05	0.04	0.09
MGO	0.80	0.83	0.75	0.74	0.82	0.78	0.70	0.93	0.79	0.83	0.91
CAO	0.93	0.88	0.90	0.78	0.84	0.84	1.01	1.19	1.13	1.14	1.28
MA2O	3.10	3.28	3.22	2.90	2.85	2.98	3.14	3.16	3.29	2.97	3.45
K2O	4.58	4.74	4.51	4.40	4.76	4.97	4.26	3.75	4.24	4.41	4.02
M2O+	0.90	0.99	1.04	1.30	1.00	1.12	0.98	1.02	0.95	1.10	0.94
M2O-	0.04	0.04	0.19	0.37	0.15	0.34	0.12	0.18	0.15	0.03	0.08
P2O5	0.19	0.24	0.17	0.15	0.18	0.23	0.10	0.20	0.20	0.22	0.16
CO2 OR LOI	0.02	0.03	0.05	0.05	0.02	0.03	0.03	0.03	0.07	0.05	0.09
TOTAL OXIDES	99.82	99.83	100.15	99.74	100.35	100.06	100.36	99.78	98.90	100.90	100.45
NORMATIVE MINERALOGY (PERCENT WEIGHT)											
QUARTZ	31.91	30.45	33.57	35.44	34.01	32.44	33.91	33.90	31.48	34.04	30.14
CORUNDUM	2.79	2.79	2.43	3.14	2.45	2.73	2.56	3.14	2.35	2.98	2.84
ORTHOCPLASE	27.41	28.40	24.93	24.55	28.34	25.77	25.39	22.51	25.45	26.10	23.88
ALBITE	24.46	28.11	27.48	25.03	24.32	25.59	24.73	27.05	28.47	25.11	29.31
ANORTHITE	3.30	2.59	3.09	2.62	3.07	2.44	4.14	4.42	4.01	3.98	4.82
EMSYATITE	2.02	2.07	1.89	1.93	2.06	1.97	2.23	2.33	1.99	2.03	2.28
FERROSILLITE	4.98	4.11	2.17	2.57	4.01	2.55	3.64	4.25	4.10	3.73	4.33
MAGNETITE	0.18	0.07	1.38	1.57	0.48	1.24	0.48	0.91	0.55	0.49	0.76
ILMENITE	0.48	0.79	0.58	0.48	0.63	0.64	0.40	0.93	0.58	0.91	0.80
APATITE	0.45	0.56	0.38	0.36	0.40	0.54	0.23	0.47	0.48	0.49	0.37
CALCITE	0.02	0.07	0.12	0.12	0.02	0.07	0.07	0.07	0.14	0.11	0.18
TRACE ELEMENTS (PPM)											
BA	290	340	350	310	340	520	380	240	320	390	380
CU	4	6	25	59	4	9	3	30	3	14	13
P	560	540	520	430	370	450	340	720	420	320	580
MO	1.6	1.2	1.0	0.9	2.1	1.0	2.2	0.4	1.0	0.9	0.9
PB	24	14	17	22	24	16	14	12	10	16	18
BB	207	204	182	184	200	154	119	196	196	201	191
BM	6.5	6.6	6.4	5.2	4.9	4.4	4.5	4.0	5.7	4.4	5.4
BR	107	107	113	83	4.9	85	115	110	105	138	120
ZN	79	73	75	96	116	115	86	219	111	100	120
U	5.8	8.7	68.7	13.9	5.7	12.3	14.3	17.1	14.8	11.6	4.0
U	4	4	4	4	4	4	4	4	4	4	4
U	34	18	34	31	12	16	45	19	31	25	25
BC	8.3	4.3	6.0	4.0	3.5	4.0	4.5	5.2	4.3	4.3	6.3
BI	105	85	85	75	74	54	79	125	121	122	117
LI	---	---	---	---	---	---	---	---	---	---	---
SB	---	---	---	---	---	---	---	---	---	---	---
CO	---	---	---	---	---	---	---	---	---	---	---
NI	---	---	---	---	---	---	---	---	---	---	---
TH	12	9	7	9	8	7	9	9	9	9	9
TR	330	270	180	190	340	390	180	280	320	210	210
AG	-.20	-.20	.50	-.20	.20	-.20	-.20	-.20	-.20	-.20	-.20

SAMPLE NO	F1-9-55	F1-9-60	F1-9-70	F1-9-74	F1-9-80	F1-9-90	F1-9-100	F1-9-105	F1-9-110	F1-9-115	F1-9-120
SiO ₂	47.75	47.21	47.79	74.17	72.94	49.90	72.10	49.29	66.27	47.35	49.43
TiO ₂	0.89	0.75	0.47	0.20	0.52	0.57	0.47	0.59	0.72	0.77	0.40
Al ₂ O ₃	15.52	15.37	15.78	14.32	13.50	14.41	13.47	14.12	13.42	14.49	14.79
Fe ₂ O ₃	0.36	0.27	0.35	0.55	1.50	0.36	0.10	0.39	0.26	0.62	0.31
FeO	3.24	3.33	3.30	0.49	1.87	3.60	2.83	3.80	4.73	3.20	3.84
MnO	0.05	0.06	0.05	0.03	0.04	0.07	0.05	0.07	0.09	0.08	0.06
MgO	1.05	1.04	1.06	0.36	0.94	1.20	0.89	1.20	1.45	1.29	1.00
CaO	1.83	2.20	1.59	0.53	0.44	1.20	0.99	1.20	2.11	1.39	1.62
Na ₂ O	3.46	3.20	2.99	3.27	2.91	3.35	2.84	3.39	3.47	3.47	2.97
K ₂ O	4.51	4.65	5.74	4.78	2.70	3.97	4.45	3.95	4.15	4.18	4.95
H ₂ O	1.05	1.04	0.78	0.78	2.21	1.13	0.80	1.24	1.18	1.15	1.00
H ₂ O ⁺	0.01	0.04	0.01	0.41	0.44	0.12	0.16	0.13	0.02	0.01	0.10
P ₂ O ₅	0.24	0.27	0.22	0.42	0.28	0.24	0.25	0.21	0.29	0.29	0.20
CO ₂ OR LOI	0.12	0.49	0.05	0.07	0.08	0.08	0.05	0.17	0.08	0.07	0.07
TOTAL OXIDES	100.10	99.94	100.34	100.78	100.55	100.97	100.58	100.87	100.32	99.00	100.50
NORMATIVE MINERALOGY (PERCENT WEIGHT)											
QUARTZ	24.28	24.71	22.14	36.44	44.46	27.84	32.40	26.21	22.11	25.39	27.29
CORUNDUM	2.53	2.84	2.39	3.79	5.95	1.87	1.78	0.92	2.49	2.44	2.38
ORTHOCASE	24.89	27.84	34.11	28.46	14.32	23.50	26.38	2.50	24.78	25.28	29.41
ALBITE	29.49	27.40	25.33	27.74	25.34	28.35	24.14	2.85	28.53	29.94	25.21
ANORTHITE	6.77	6.25	6.23	0.78	0.01	7.54	4.43	9.19	8.14	4.72	4.34
EMSTATITE	2.62	2.60	2.65	0.78	2.44	3.00	2.23	3.01	3.45	3.24	2.51
FERROSILLITE	4.34	4.80	4.76	0.48	1.44	5.48	4.25	5.82	7.50	5.54	4.77
MAGNETITE	0.53	0.40	0.51	0.80	2.22	0.52	0.44	0.57	0.38	0.92	0.45
ILMENITE	1.71	1.44	1.28	0.38	0.99	1.09	0.90	1.11	1.16	1.48	1.15
APATITE	0.59	0.41	0.49	0.94	0.44	0.58	0.54	0.47	0.44	0.69	0.47
CALCITE	0.25	1.11	0.11	0.14	0.19	0.18	0.11	0.37	0.18	0.14	0.14
TRACE ELEMENTS (PPM)											
BA	750	950	1600	390	200	830	800	710	520	620	560
CU	9	8	12	121	245	7	9	8	29	10	7
F	490	520	480	310	460	520	430	560	460	560	520
MO	1.3	2.2	2.1	2.2	4.1	1.4	3.7	2.0	1.8	2.0	1.2
PR	30	40	46	48	30	20	40	14	20	42	18
SR	144	140	166	128	124	165	151	145	128	199	199
SM	4.3	4.7	4.7	5.9	6.2	5.4	4.7	5.1	5.1	5.2	5.0
BR	120	75	135	74	105	170	150	165	125	120	120
ZN	66	254	200	124	340	80	238	73	95	248	41
U	3.2	3.2	11.8	113.0	748.0	4.5	3.4	3.0	2.7	3.4	3.4
N	4	4	4	4	4	4	4	4	4	4	4
B	19	9	8	37	98	17	19	15	4	35	18
PC	3.5	3.7	1.7	5.5	4.2	3.9	3.4	3.4	3.9	3.5	4.0
BI	---	---	---	---	---	---	---	---	---	---	---
LI	84	49	72	19	44	82	54	72	89	90	49
SB	---	---	---	---	---	---	---	---	---	---	---
CO	---	---	---	---	---	---	---	---	---	---	---
MI	---	---	---	---	---	---	---	---	---	---	---
TH	10	10	10	2	5	10	9	11	14	15	9
ZR	300	200	225	240	320	240	320	240	505	315	550
AG	20	20	20	1.20	1.90	20	1.30	20	20	20	20

SAMPLE NO	25.5	71.0	10-34.3	10-51.8	10-56.5	11-21.9	11-50.9	10-61.7	11-18.3	11-20	11-27.3	11-33
SI02	70.77	71.91	70.91	69.05	68.57	72.72	73.90	68.02	68.17	70.29	69.47	69.82
TI02	0.75	0.70	0.72	0.49	0.67	0.59	0.74	0.75	0.71	0.78	0.76	0.76
AL2O3	12.96	13.39	13.28	14.36	13.77	13.77	12.08	14.17	14.98	14.20	14.10	14.50
FE2O3	1.22	0.57	0.95	0.48	0.26	0.25	0.11	0.24	0.53	0.47	0.38	0.35
FE0	3.18	3.63	3.34	3.38	3.01	2.46	3.91	4.14	3.85	4.12	3.71	3.37
MNO	0.09	0.04	0.07	0.05	0.04	0.07	0.07	0.07	0.09	0.09	0.07	0.07
MGO	1.32	1.07	1.15	1.17	1.56	0.92	1.13	2.45	1.25	1.26	1.15	1.10
CAD	0.99	1.90	1.34	0.78	1.00	0.96	1.52	2.18	1.55	1.31	1.86	2.00
MA2O	3.54	2.67	2.84	3.38	3.30	2.81	2.48	3.13	3.15	2.98	3.00	3.16
MA2O	3.43	3.93	3.98	4.01	4.62	4.78	3.52	3.44	4.43	3.85	3.90	3.95
M2O+	1.30	0.79	1.13	1.39	0.88	1.19	0.76	0.97	1.16	1.16	0.81	0.80
M2O-	0.26	0.07	0.23	0.00	0.02	0.21	-0.1	0.02	0.03	-0.1	-0.1	0.08
F2O5	0.24	0.20	0.25	0.18	0.17	0.07	0.19	0.17	0.23	0.23	0.20	0.20
CO2 OR LOI	0.19	0.05	0.05	0.07	0.10	0.07	0.11	0.09	0.25	0.09	0.07	0.19
TOTAL OXIDES	100.24	100.94	100.24	98.99	98.94	100.97	100.51	99.84	100.31	100.82	99.47	100.35
NORMATIVE MINERALOGY (PERCENT WEIGHT)												
QUARTZ	32.88	34.03	33.84	29.95	26.28	33.86	39.28	26.57	26.65	31.78	30.12	29.50
COORUNDUM	2.66	1.89	2.60	3.73	3.04	2.77	2.12	1.98	3.29	3.47	2.23	2.31
ORTHOCASE	20.57	23.24	23.82	24.25	27.87	28.42	20.82	20.54	26.43	22.81	23.39	23.50
ALBAITE	30.37	22.59	24.31	29.23	20.82	23.89	21.04	26.72	24.89	25.31	25.74	26.80
ANDRINITE	2.19	7.75	4.75	2.37	3.28	3.29	5.68	9.31	4.79	4.57	7.60	7.52
ENSTATAITE	3.33	2.66	2.87	2.99	3.96	3.28	2.82	6.18	3.14	3.13	2.88	2.76
FERROSILITE	3.79	5.13	4.36	4.86	4.40	3.47	6.01	6.34	5.64	6.06	5.41	4.78
MAGNETITE	1.79	0.83	1.39	0.71	0.39	0.35	0.16	0.35	0.78	0.68	0.56	0.51
ILMENITE	1.44	1.31	1.36	1.34	1.30	1.11	1.39	1.44	1.36	1.49	1.46	1.45
APATITE	0.54	0.46	0.59	0.43	0.40	0.40	0.44	0.40	0.49	0.51	0.47	0.44
CALCITE	0.44	0.11	0.12	0.14	0.23	0.14	0.23	0.18	0.55	0.18	0.14	0.44
TRACE ELEMENTS (PPM)												
BA	378	665	670	550	750	710	470	440	620	550	660	660
CU	410	410	600	580	520	460	540	10	20	11	16	14
CF	1.1	2.1	2.8	3.5	2.7	1.5	2.9	2.5	2.3	1.9	2.3	2.3
MO	42	22	20	26	165	25	20	12	43	22	22	15
PPB	144	146	183	184	165	186	165	139	185	192	168	138
SM	6.6	4.3	5.0	4.7	4.8	4.0	3.7	5.4	3.8	4.1	4.3	4.8
SR	112	98	98	120	153	112	100	113	113	113	138	141
ZN	276	68	103	95	125	106	102	82	149	126	82	86
U	11.3	3.9	11.2	4.9	2.5	35.6	3.0	2.8	7.2	3.3	3.7	3.5
W	20	32	4	-4	-4	-4	-4	-4	-4	-4	-4	-4
RE	24	170	10	8	6	19	12	6	31	30	12	13
BE	2.7	1.4	3.0	1.5	2.6	3.6	2.9	2.0	2.8	3.7	3.7	4.4
BI	0.02	0.06	0.06	0.06	0.05	0.08	0.06	0.06	0.09	0.07	0.08	0.08
SB	96	80	85	92	75	87	81	81	104	117	83	69
CO	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3
NI	10	7	7	3	7	9	8	8	11	10	10	7
TH	7	13	13	10	14	14	12	14	17	11	14	15
TM	12	11	11	11	9	10	11	12	11	12	10	13
ZK	---	---	---	440	145	260	190	170	270	480	315	200
ZG	---	---	---	---	---	---	-10	---	---	---	-10	-20

SAMPLE NO	7.62	19.8	23.0	32.0	32.3	32.3A	33.0	38.0	53.6	74.69	86.0	89.0
SiO ₂	69.09	69.70	69.33	68.78	67.47	67.72	69.20	69.98	66.14	69.14	66.66	67.95
Y ₂ O ₃	0.70	0.65	0.75	0.60	0.74	0.84	0.80	0.42	0.80	0.60	0.64	0.84
Al ₂ O ₃	13.75	14.29	13.64	14.51	14.96	14.00	14.10	13.86	15.00	15.37	16.26	15.38
Fe ₂ O ₃	0.77	0.48	0.97	0.64	0.88	1.56	1.06	0.72	0.37	0.48	0.47	0.53
FeO	4.13	3.66	3.25	3.07	3.60	3.40	4.05	3.47	3.74	3.47	3.16	4.38
MnO	0.08	0.04	0.07	0.06	0.04	0.07	0.08	0.06	0.07	0.06	0.05	0.07
MgO	1.34	1.22	1.26	1.00	1.27	1.21	1.36	1.12	1.19	1.07	0.98	1.34
CaO	1.95	1.43	1.40	1.15	1.07	1.07	1.39	1.21	2.24	2.42	2.62	2.30
Na ₂ O	3.04	2.95	3.16	3.17	3.51	3.08	3.05	2.93	3.45	3.45	4.10	3.57
K ₂ O	3.75	3.24	3.42	3.09	4.02	4.47	3.91	3.80	4.17	3.52	3.92	3.21
H ₂ O ⁺	0.79	1.12	1.20	0.79	1.46	1.18	1.21	1.79	0.91	0.53	0.69	0.03
H ₂ O ⁻	0.16	0.24	0.13	0.18	0.27	0.17	0.14	0.01	0.13	0.14	0.09	0.03
P ₂ O ₅	0.25	0.25	0.23	0.23	0.23	0.23	0.23	0.23	0.24	0.23	0.22	0.24
CO ₂ OR LOI	0.07	0.09	0.23	0.12	0.07	0.12	0.05	0.09	0.09	0.09	0.17	0.14
TOTAL OXIDES	99.89	99.38	99.36	99.39	99.61	99.12	100.65	99.89	98.54	100.73	100.03	100.77

NORMATIVE MINERALOGY (PERCENT WEIGHT)

QUARTZ	29.25	34.48	32.89	26.38	26.65	28.59	30.17	33.46	22.61	27.24	19.97	26.22
CORDIERITE	1.87	4.20	3.35	2.55	3.42	3.05	3.06	3.21	1.54	2.21	1.42	2.74
ORTHOCASE	22.43	19.51	20.66	30.61	24.31	27.00	23.25	22.87	25.26	20.76	23.37	18.95
ALBITE	24.18	25.48	27.21	27.27	30.36	26.58	25.92	25.99	29.96	29.19	34.88	30.16
ANORTHITE	7.74	5.07	4.09	3.51	3.57	3.18	5.05	4.14	9.29	10.06	10.63	9.05
ENSTATITE	3.37	3.08	3.18	2.53	3.23	3.08	3.39	2.82	3.02	2.67	2.46	3.34
FERROSILITE	5.99	5.47	4.69	4.26	4.88	3.78	5.41	4.94	5.48	5.35	4.47	6.34
MAGNETITE	1.13	0.71	1.44	0.94	1.30	2.32	1.55	1.07	0.55	0.70	0.69	0.77
ILMENITE	1.33	1.24	1.45	1.16	1.42	1.61	1.53	1.20	1.56	1.14	1.23	1.58
APATITE	0.59	0.57	0.52	0.52	0.52	0.52	0.56	0.52	0.55	0.51	0.51	0.53
CALCITE	0.14	0.19	0.53	0.28	0.14	0.28	0.12	0.19	0.19	0.18	0.37	0.30

TRACE ELEMENTS (PPM)

BA	440	480	620	790	550	760	550	540	680	590	460	480
CU	9	9	9	117	276	195	120	10	12	9	11	13
F	400	460	490	320	430	460	520	340	460	460	400	430
MO	1.4	1.1	1.9	1.0	2.4	1.7	2.1	2.2	1.4	1.5	1.0	1.4
PB	34	36	28	18	18	26	14	40	20	26	24	12
RB	166	150	167	178	176	196	196	182	156	115	103	127
SM	5.2	4.4	3.9	3.4	4.5	4.3	4.6	4.1	5.0	4.3	3.9	4.5
SK	140	116	122	150	120	150	123	105	167	140	145	150
ZN	133	307	135	120	205	150	136	230	73	73	65	82
U	4.3	4.2	7.8	5.8	6.2	4.5	4.7	6.7	3.1	6.6	3.6	4.7
W	-4	32	12	8	-4	-4	12	24	-4	-4	24	16
B	14	23	18	16	99	21	29	27	14	8	6	10
BE	3.6	2.3	2.3	2.0	3.1	3.0	3.0	2.4	1.8	2.3	1.5	3.1
BI	0.06	0.08	0.06	0.80	0.31	0.07	0.10	0.06	0.08	0.08	0.07	0.08
LI	100	130	133	90	105	102	108	109	77	77	76	93
SB	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3
CO	10	6	5	7	7	9	8	7	8	9	3	4
NI	12	7	18	6	16	8	15	2	15	12	8	12
TM	11	13	12	7	12	12	14	9	11	11	10	10

SAMPLE NO	10-42.1	10-43.2	10-48.8	10-50.2	10-60.0	10-61.5	11-13.2	11-34.2	11-44.8	11-63.5	11-64.2	11-67
SiO ₂	70.91	67.82	66.65	68.41	67.21	67.32	68.70	69.89	67.38	67.38	69.07	67.67
TiO ₂	0.47	0.77	0.75	0.63	0.79	0.74	0.66	0.74	1.17	1.17	1.00	0.93
Al ₂ O ₃	13.47	15.10	15.05	14.64	14.31	14.68	15.07	14.30	15.51	15.51	14.64	15.55
FeO	0.40	0.38	0.43	0.66	0.33	0.27	0.61	0.22	0.23	0.23	0.35	0.18
Fe ₂ O ₃	3.48	4.35	3.58	3.13	4.25	3.99	3.58	4.05	4.64	4.64	3.93	3.84
MnO	0.06	0.08	0.07	0.05	0.08	0.07	0.08	0.08	0.08	0.08	0.08	0.07
MgO	1.07	1.34	1.17	1.06	1.27	1.27	1.09	1.24	1.41	1.41	1.21	1.19
CaO	1.57	1.60	1.45	1.56	2.00	1.93	1.72	1.88	1.87	1.87	1.91	1.92
Na ₂ O	2.77	3.18	3.28	3.70	3.05	3.22	3.25	3.04	3.07	3.20	3.06	3.21
K ₂ O	4.22	4.00	4.75	3.84	4.04	4.85	3.96	3.61	4.14	4.04	4.05	4.05
H ₂ O ⁺	0.86	1.06	1.04	1.23	1.21	0.83	0.98	0.84	0.96	0.96	0.85	0.86
H ₂ O ⁻	0.03	0.16	0.08	0.02	0.02	0.09	0.08	0.13	0.13	0.13	0.12	0.08
P ₂ O ₅	0.21	0.25	0.17	0.17	0.21	0.74	0.21	0.23	0.24	0.20	0.20	0.07
CO ₂ OR LOI	0.24	0.16	0.31	0.07	0.21	0.10	0.05	0.14	0.05	0.14	0.14	0.07
TOTAL OXIDES	100.36	100.25	98.78	99.39	98.98	100.30	100.04	100.41	100.67	100.97	100.60	99.82
NOMINATIVE MINERALOGY (PERCENT WEIGHT)												
QUARTZ	32.72	27.00	24.10	26.48	26.74	23.87	28.05	30.94	26.68	26.68	28.50	26.14
CRUMMUM	2.79	3.63	3.07	2.14	2.30	2.86	2.95	2.87	2.96	2.96	2.94	3.04
ORTHOCASE	25.04	23.90	28.77	23.15	24.40	28.83	23.62	21.43	24.47	24.47	23.94	24.24
ALBITE	23.48	27.17	28.42	31.91	26.33	27.35	27.80	25.88	27.12	27.12	26.00	27.40
ANORTHITE	4.93	5.34	4.22	6.37	7.52	4.14	6.93	6.98	7.68	7.68	7.39	8.00
ENSTHATE	2.68	3.37	2.98	2.69	3.24	3.19	2.74	3.11	3.49	3.49	3.03	2.97
FERROSILITE	3.44	4.62	5.22	4.30	6.53	6.03	5.15	6.22	6.35	6.35	5.45	5.58
MAGNETITE	0.58	0.56	0.64	0.98	0.49	0.39	0.89	0.32	0.33	0.33	0.51	0.26
ILMENITE	1.28	1.46	1.44	1.24	1.52	1.40	1.27	1.40	2.23	2.23	1.91	1.77
APATITE	0.49	0.59	0.40	0.40	0.48	1.71	0.47	0.51	0.44	0.44	0.44	0.45
CALCITE	0.55	0.37	0.72	0.14	0.47	0.23	0.12	0.34	0.11	0.11	0.30	0.14
TRACE ELEMENTS (PPM)												
BA	420	610	740	510	420	590	650	520	640	595	600	460
CU	10	9	5	2	5	8	46	16	24	15	12	13
F	510	640	520	500	600	600	510	520	430	640	520	460
NO	4.7	3.7	1.8	3.4	2.9	2.9	2.0	3.0	3.1	1.7	2.2	2.1
PB	36	36	36	30	17	20	32	20	22	24	20	16
KB	184	215	207	182	165	155	173	170	137	186	178	164
SW	5.0	5.5	5.0	5.0	5.5	5.5	3.9	4.9	3.2	3.4	3.7	3.6
SR	142	142	142	142	132	135	130	140	130	130	120	155
ZN	195	264	120	102	214	94	104	84	126	111	81	174
U	6.0	6.9	3.5	4.0	3.3	4.5	6.9	3.6	7.5	4.2	3.9	3.8
W	-4	-4	20	-4	-4	-4	-4	-4	-4	-4	-4	-4
B	6	9	10	11	3	3	22	14	34	12	7	7
BE	1.7	2.8	2.8	1.9	2.3	2.8	4.6	2.7	4.0	4.3	5.6	5.3
BI	0.06	0.07	0.05	0.06	0.06	0.06	0.08	0.10	0.09	0.06	0.10	0.14
LI	84	109	91	81	106	84	106	84	75	95	89	78
SB	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3
CO	7	6	6	9	8	9	8	7	8	9	7	4
NI	14	14	11	13	16	15	14	18	13	13	10	10
YH	10	13	10	10	12	11	11	12	10	13	13	10
ZK	230	185	230	185	230	150	170	250	370	220	200	310
AG	---	---	---	---	---	---	40	20	40	10	10	10

SAMPLE NO	25.7	41.0	80-10-24	80-10-46	83.0	39.8	F1-1-35	F1-1-68.9	F1-7-5	F1-7-25	F1-7-40	F1-7-50
SiO2	70.11	67.55	67.89	61.12	66.62	48.11	71.98	69.51	71.11	68.89	69.02	67.33
TiO2	0.55	0.72	0.38	1.00	0.65	0.89	0.25	0.60	0.35	0.53	0.50	0.57
Al2O3	13.97	15.36	15.48	17.43	15.88	13.72	13.33	14.73	14.85	16.29	14.67	14.82
Fe2O3	0.55	0.42	0.88	0.63	0.51	1.20	0.05	0.45	0.85	0.18	0.22	0.54
FeO	3.24	3.66	2.57	5.14	3.74	3.57	2.88	3.84	2.79	2.90	3.44	3.29
MnO	0.04	0.07	0.04	0.10	0.07	0.06	0.07	0.07	0.06	0.04	0.06	0.07
MgO	1.15	1.19	0.74	1.67	1.25	1.40	0.84	1.35	1.00	0.96	1.17	1.23
CaO	1.15	1.73	1.24	1.87	1.60	0.82	1.28	1.80	1.31	1.60	2.12	1.97
K2O	3.52	3.80	3.25	4.03	3.13	3.75	3.22	3.43	3.06	3.45	3.32	2.84
Na2O	3.75	3.91	4.78	4.72	5.50	3.73	3.99	3.60	4.27	4.47	3.97	4.40
P2O5	1.53	1.24	0.99	1.32	1.01	1.15	1.09	0.88	1.05	0.98	1.27	1.70
CO2 OR LOI	0.08	0.00	0.10	0.09	0.06	0.00	0.04	0.10	0.11	0.13	0.09	0.08
TOTAL OXIDES	100.17	100.45	98.59	99.71	100.68	98.55	99.60	100.81	100.61	100.74	100.57	99.72
NORMATIVE MINERALOGY (PERCENT WEIGHT)												
QUARTZ	30.84	25.04	26.73	11.60	21.46	27.81	33.94	28.62	31.39	25.82	27.30	28.74
ORTHOCLASTE	3.24	3.41	3.38	3.76	3.39	2.90	2.78	3.02	3.26	3.59	2.30	3.92
ALBITE	22.51	23.28	29.01	28.36	32.68	22.66	23.92	21.28	25.40	26.49	23.62	26.59
ANORTHITE	30.14	32.35	28.21	34.70	24.52	32.59	27.60	29.08	26.04	29.31	28.24	24.55
ENSTATITE	2.54	3.64	4.66	5.55	3.84	1.92	2.75	5.98	5.35	4.01	7.31	4.27
FERROSILITE	2.88	2.97	1.87	4.23	3.13	3.56	2.10	3.37	2.51	2.40	2.94	3.10
MAGNETITE	4.79	5.34	3.51	7.57	5.53	4.64	5.01	5.83	4.41	5.84	5.84	4.87
ILMENITE	0.81	0.41	1.31	0.93	1.22	1.35	0.07	0.65	0.66	0.26	0.32	0.80
APATITE	1.04	1.36	0.74	1.93	0.54	0.57	0.48	1.14	0.65	1.01	0.96	1.11
CALCITE	0.57	0.56	0.34	0.45	0.94	0.23	0.86	0.34	0.40	0.58	0.70	0.47
TRACE ELEMENTS (PPM)	0.58	1.22	0.23	0.93	0.94	0.23	0.86	0.34	0.02	0.11	0.71	1.56
BA	395	560	1600	760	945	500	340	450	450	740	550	640
CU	7	7	2	30	22	12	3	5	12	10	18	22
F	430	460	410	670	490	430	540	580	360	450	490	460
MO	2.1	2.6	1.3	2.4	1.7	1.7	2.3	1.9	2.6	1.9	2.8	5.7
PB	68	54	42	320	186	72	18	180	40	36	34	38
RB	140	149	218	200	167	147	191	150	165	154	147	180
SN	4.1	4.1	4.0	5.1	4.5	4.4	6.7	4.2	4.3	4.2	5.1	7.2
SR	100	160	152	163	145	92	83	130	130	130	132	110
ZN	216	184	89	720	435	355	89	75	246	75	160	570
U	2.6	5.2	2.4	3.7	3.4	10.3	4.5	3.6	2.3	3.1	4.8	2.4
W	44	8	-4	-4	4	4	-4	4	-4	-4	-4	-4
B	70	25	3	8	33	21	23	49	38	21	21	54
BE	2.4	1.4	1.2	3.9	1.0	1.9	5.3	2.8	2.1	3.3	3.7	3.7
BI	0.06	0.05	0.05	0.06	0.09	0.05	-	-	-	-	-	-
LI	115	115	57	110	104	109	89	94	72	108	96	76
SB	-3	-3	-3	-3	-3	-3	-	-	-	-	-	-
CO	4	7	2	9	5	7	-	-	-	-	-	-
NI	5	11	7	15	18	13	-	-	-	-	-	-
TH	11	14	6	12	8	10	10	11	9	11	11	11
ZK	11	14	6	12	8	10	10	11	9	11	11	11
AG	145	145	145	275	145	145	380	240	195	170	250	240
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SAMPLE NO	82.0	80-10-11	80-10-22	80-11-5	80-11-27	80-11-56	35.5A	56.00	80-10-13	80-10-14	80-10-30	80-11-6
SiO2	70.88	48.23	46.64	48.83	48.20	46.60	68.54	66.46	48.91	49.47	48.56	49.74
TiO2	0.87	0.47	0.45	0.78	0.93	0.98	0.67	0.44	0.66	0.80	0.55	0.88
Al2O3	12.51	15.36	16.31	14.76	15.55	15.74	15.12	16.12	14.12	14.13	14.71	14.34
Fe2O3	1.16	0.71	0.55	0.92	0.86	0.69	0.81	0.58	0.35	0.73	0.47	1.34
FeO	3.45	3.29	3.27	3.92	3.89	4.72	3.67	3.81	3.33	3.98	3.34	2.96
MnO	0.07	0.07	0.07	0.09	0.08	0.09	0.07	0.07	0.06	0.07	0.07	0.07
MgO	1.11	1.13	1.09	1.40	1.25	1.52	1.38	1.18	1.06	1.35	1.11	1.05
CaO	2.28	2.32	2.01	1.49	2.37	2.33	1.12	2.30	1.91	1.84	1.79	1.65
Na2O	3.10	3.77	3.67	3.41	3.94	3.52	4.13	3.66	3.26	3.14	3.38	2.83
K2O	2.68	3.25	4.30	3.22	2.16	2.50	2.74	4.10	4.14	3.30	3.58	3.54
H2O+	0.81	1.21	0.98	1.44	1.17	1.20	1.06	0.71	1.04	1.18	1.01	0.99
H2O-	0.03	0.11	0.01	0.09	0.15	-0.1	0.23	0.11	0.07	0.10	0.12	-0.1
P2O5	0.19	0.20	0.19	0.21	0.21	0.18	0.23	0.23	0.17	0.21	0.19	0.23
CO2 OR LOI	0.09	0.05	0.14	0.04	0.09	0.09	0.09	0.07	0.05	0.05	0.10	0.05
TOTAL OXIDES	99.23	100.57	99.88	100.60	100.65	100.15	99.86	99.84	99.13	100.35	98.98	99.46
NORMATIVE MINERALOGY (PERCENT WEIGHT)												
QUARTZ	35.58	26.12	21.91	30.02	28.78	27.38	28.66	21.59	27.41	31.13	29.02	34.51
CORDIERITE	1.00	2.24	3.72	3.56	3.14	3.70	4.17	2.24	1.36	2.67	2.80	3.58
ORTHOCASE	16.12	19.38	25.74	19.18	12.81	14.95	16.39	24.45	24.94	19.65	21.60	21.28
ALBITE	26.59	32.07	31.42	29.06	33.50	30.02	35.19	31.21	28.15	26.83	29.16	24.25
ANORTHITE	9.79	9.98	9.98	9.98	10.02	9.93	9.98	9.98	8.23	7.58	7.11	6.48
ENSTATHITE	2.79	2.84	2.75	3.50	3.14	3.83	3.49	2.95	2.69	3.40	2.80	2.63
FERRUGILITE	4.14	4.50	4.64	5.35	5.25	6.47	5.15	5.24	4.93	5.57	5.06	3.72
MAGNETITE	1.71	1.04	0.81	1.35	0.96	1.03	1.19	0.85	1.07	0.70	0.70	1.29
ILMENITE	1.68	1.28	1.23	1.50	0.96	1.88	1.29	1.23	1.28	1.53	1.07	1.98
APATITE	0.43	0.44	0.42	0.47	0.47	0.42	0.54	0.54	0.38	0.47	0.45	0.52
CALCITE	0.19	0.12	0.30	0.07	0.18	0.18	0.19	0.14	0.12	0.12	0.23	0.12
TRACE ELEMENTS (PPM)												
BA	280	540	730	490	470	340	450	650	600	500	470	690
CU	3	35	44	17	9	28	23	8	13	24	5	19
F	520	540	580	620	420	450	430	430	560	670	600	640
MO	1.5	2.3	2.6	3.0	3.4	4.5	2.7	1.8	1.5	3.4	2.3	2.5
PB	20	32	30	42	16	18	42	22	22	30	52	17
KB	125	163	205	175	133	184	176	160	180	187	188	165
SN	3.9	4.6	4.6	5.1	4.7	4.1	4.1	4.5	5.2	5.6	4.8	4.5
SR	130	195	165	135	117	135	140	180	170	140	118	115
CR	70	100	91	180	96	215	390	72	111	131	204	184
ZN	3.5	6.5	20.6	10.7	23.8	4.4	22.4	4.0	4.9	6.9	6.5	5.4
U	-4	-4	-4	-4	-4	-4	-4	68	-4	-4	-4	-4
M	7100	10	10	40	24	22	27	13	12	12	16	22
BE	1.2	3.0	2.9	4.2	3.8	4.5	1.8	2.5	2.9	2.9	2.1	3.0
BI	0.17	0.06	0.06	0.06	0.09	0.08	0.05	0.13	0.06	0.08	0.06	0.06
LI	88	84	95	96	84	113	104	94	96	96	94	88
SB	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3	-3
CO	6	3	3	8	12	11	7	10	9	3	5	9
CD	4	4	4	11	13	20	4	14	4	15	10	12
NI	8	13	11	11	15	20	4	14	9	15	10	12
IN	31	10	12	12	15	11	8	10	10	13	11	11
ZR	---	280	240	330	540	260	---	---	440	265	---	530
AS	---	---	---	10	10	60	---	---	---	---	---	20

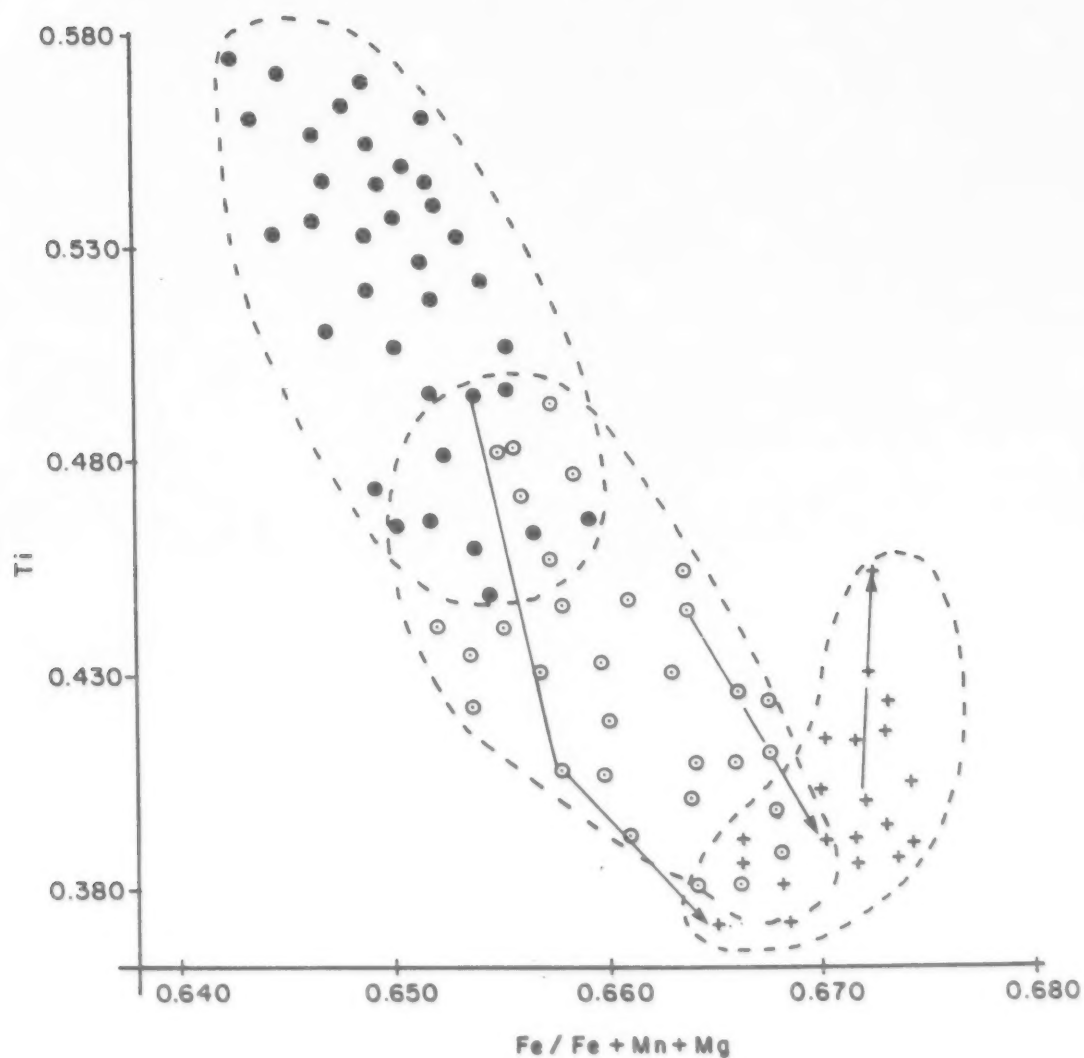


Figure 5-9. Ti vs $Fe/Fe+Mn+Mg$ plot showing the composition of biotite from the unaltered granodiorite (.), altered granodiorite (o) and mineralized granodiorite (+) of the Millet Brook area. Sample points joined by solid line represent composition of biotite within a single thin section.

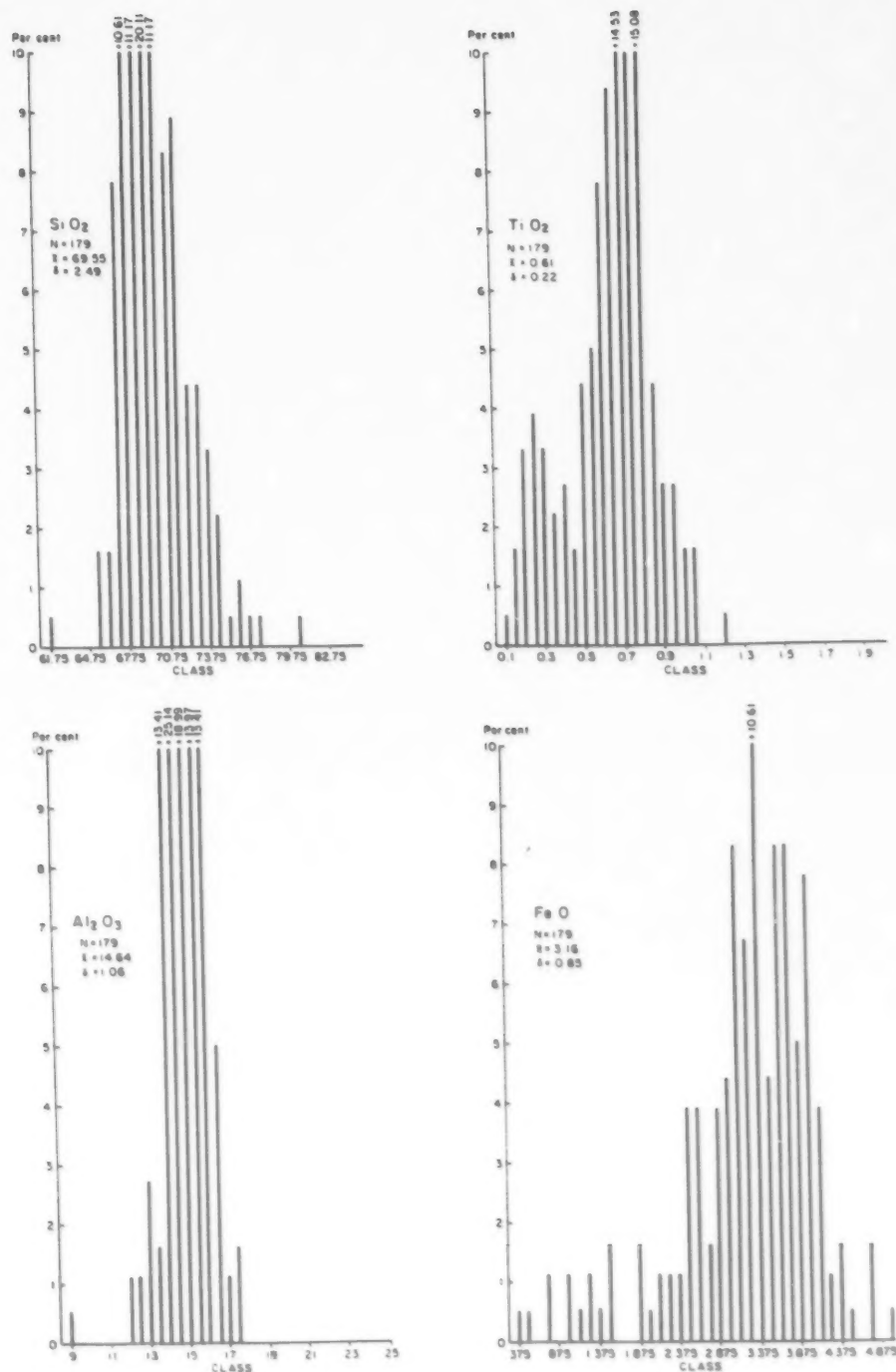


Figure 5-10. Histograms showing the frequency distribution of SiO₂, TiO₂, Al₂O₃ and FeO from the granodiorite of the Millet Brook area (analytical data from Table 5-2).

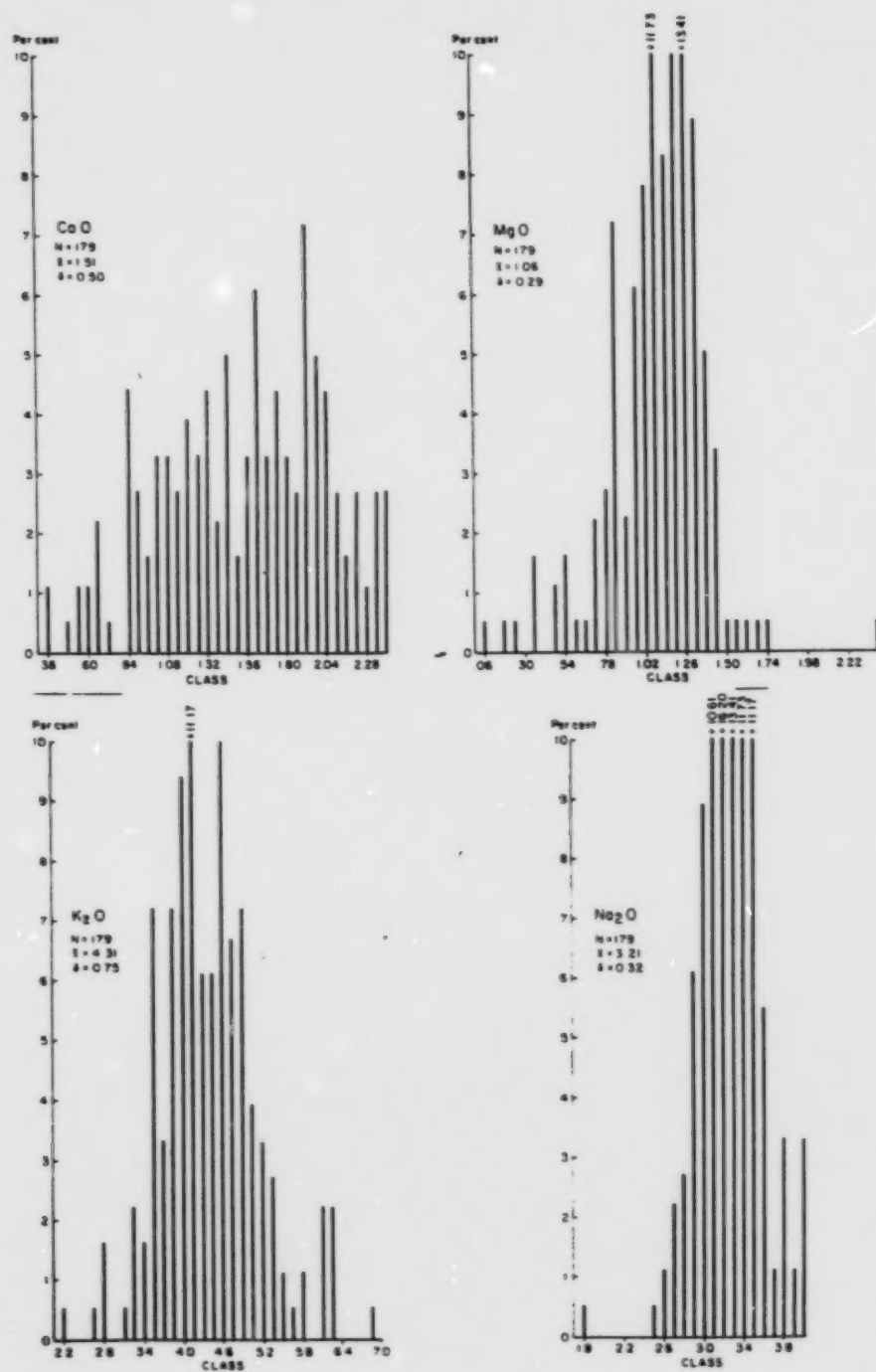


Figure 5-11. Histograms showing the frequency distribution of CaO, MgO, K₂O and Na₂O from the granodiorites of the Millet Brook area (analytical data from Table 5-2).

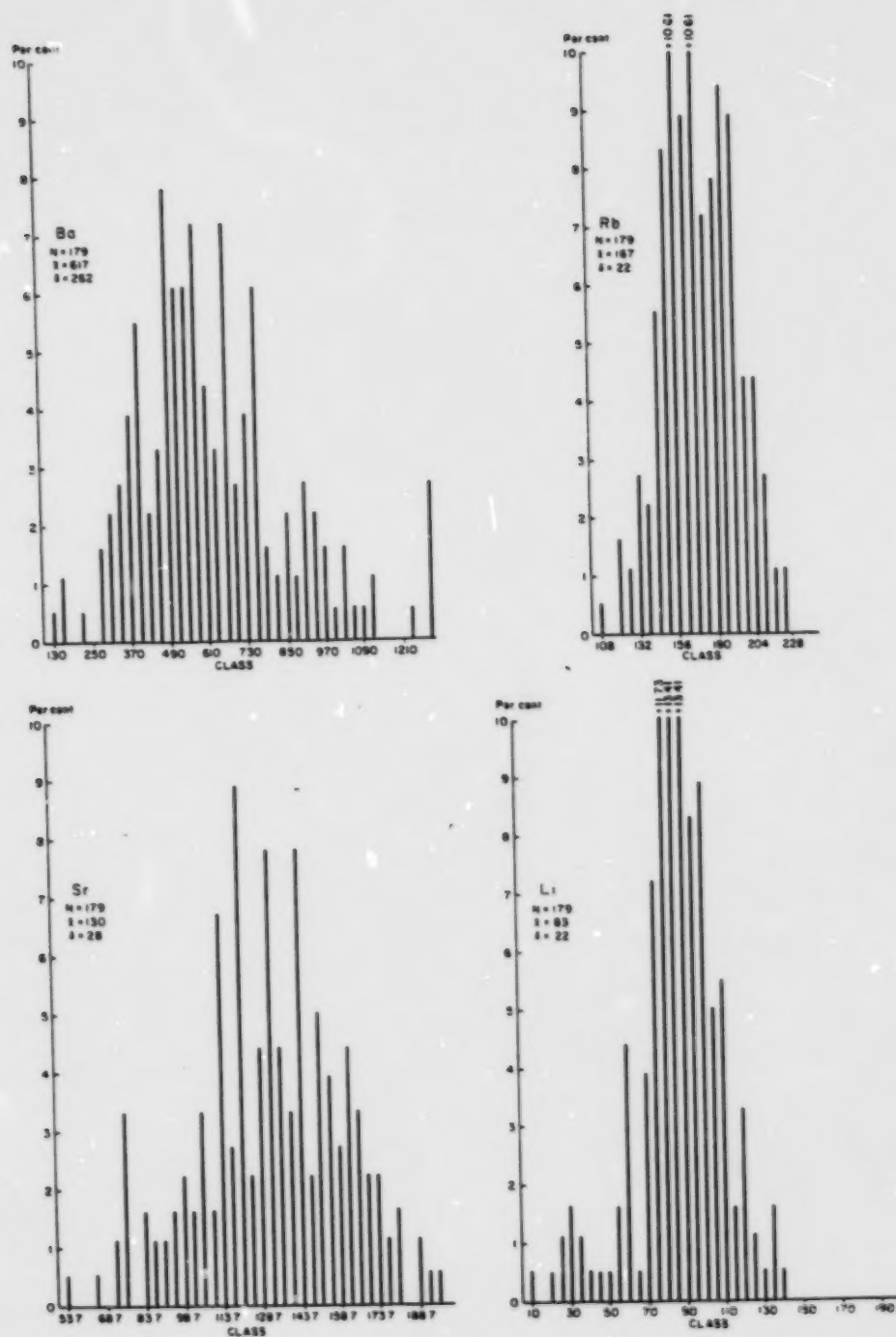


Figure 5-12. Histograms showing the frequency distribution of Ba, Rb, Sr and Li from the granodiorites of the Millet Brook area (analytical data from Table 5-2).

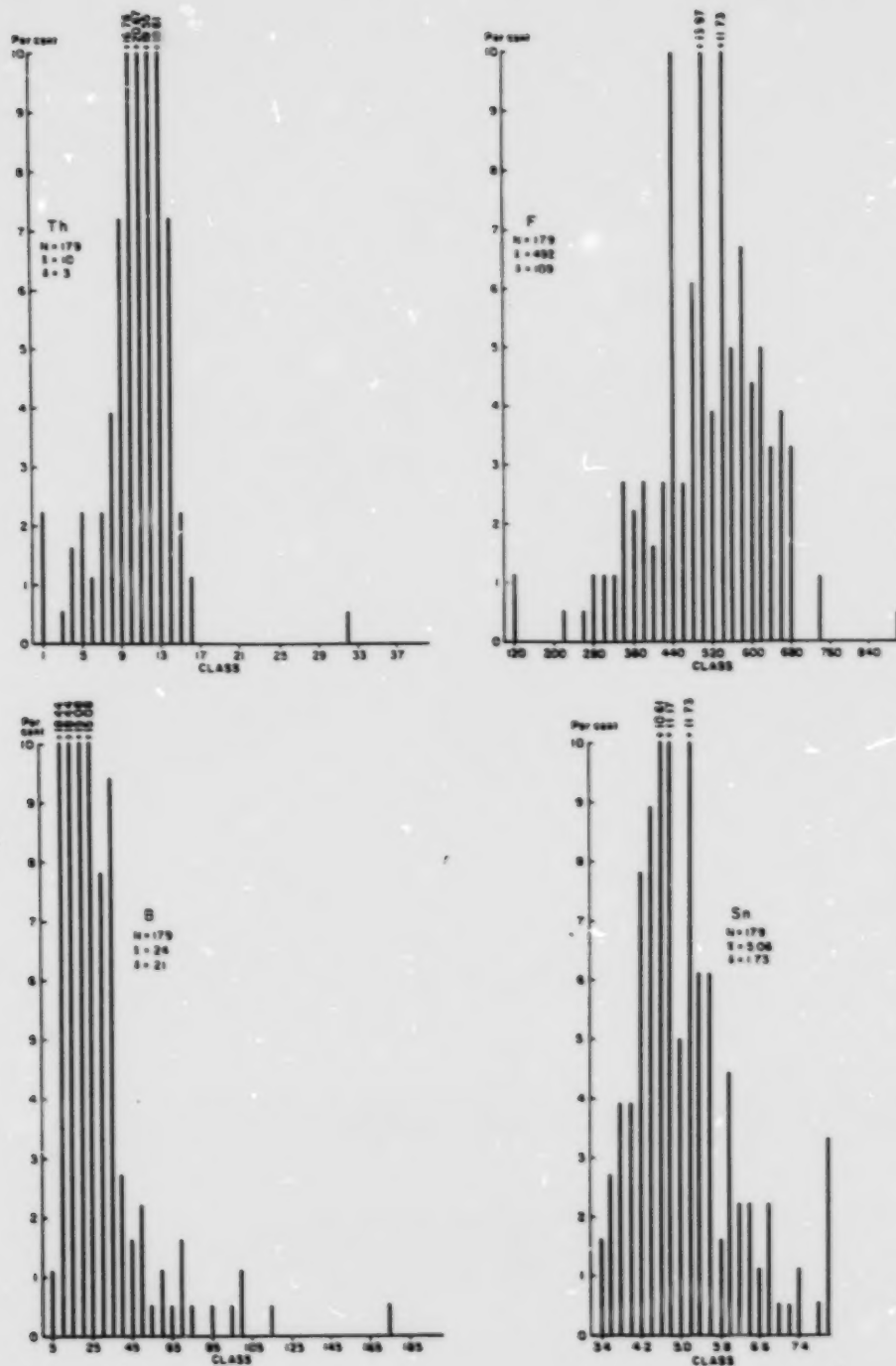


Figure 5-13. Histograms showing the distribution of Th, F, B and Sn from the granodiorite of the Millet Brook area (analytical data from Table 5-2).

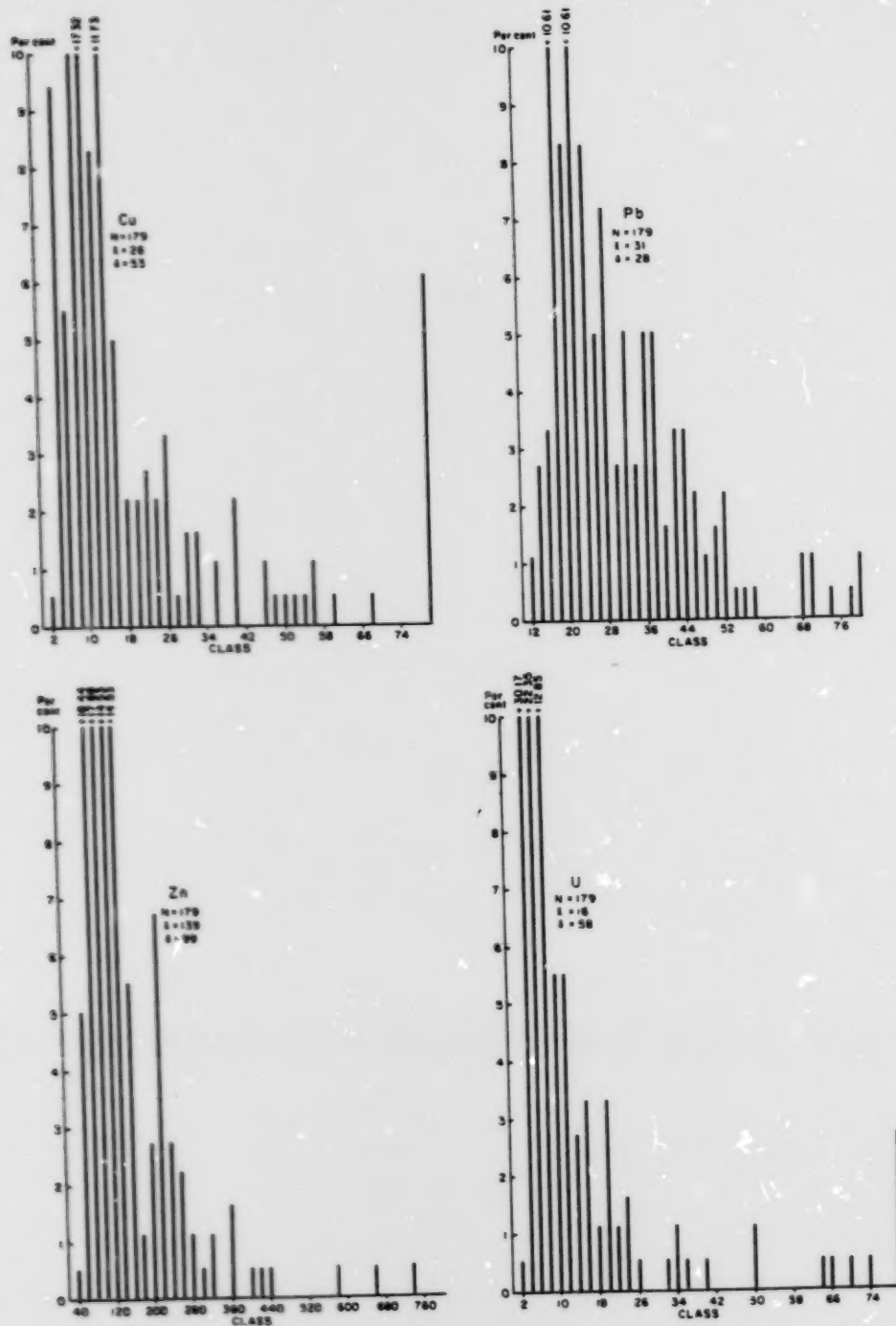


Figure 5-14. Histograms showing the distribution of Cu, Pb, Zn and U from the granodiorites of the Millet Brook area (analytical data from Table 5-2).

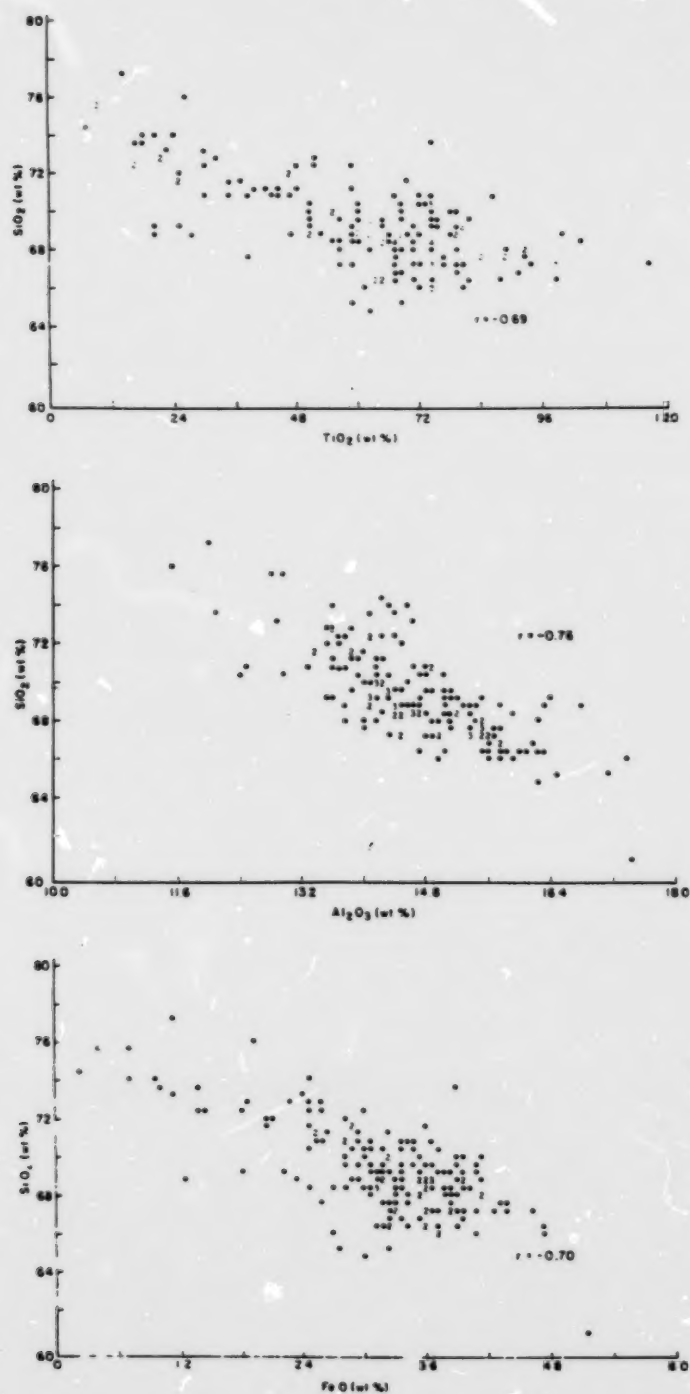


Figure S-15. Silica variation diagram for the granodiorites of the Millet Brook area.

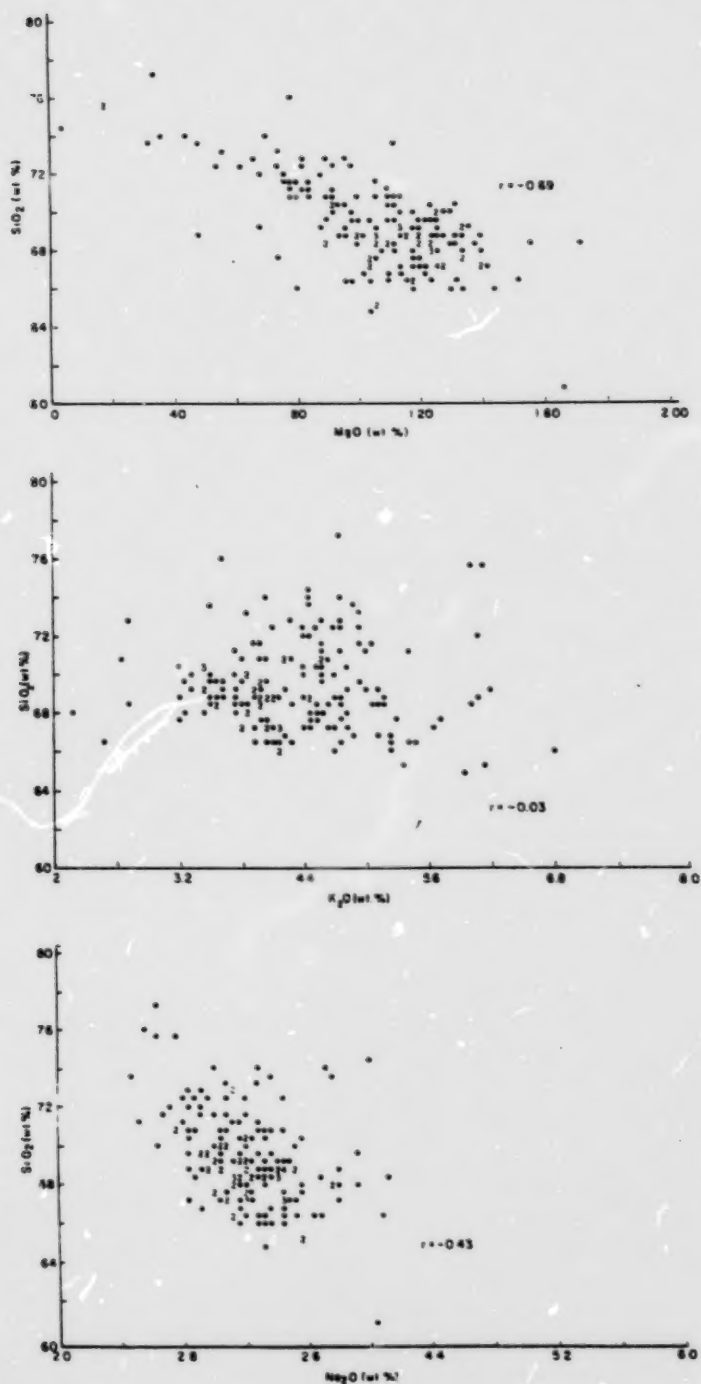


Figure 5-16. Silica variation diagram for the granodiorites of the Millet Brook area.

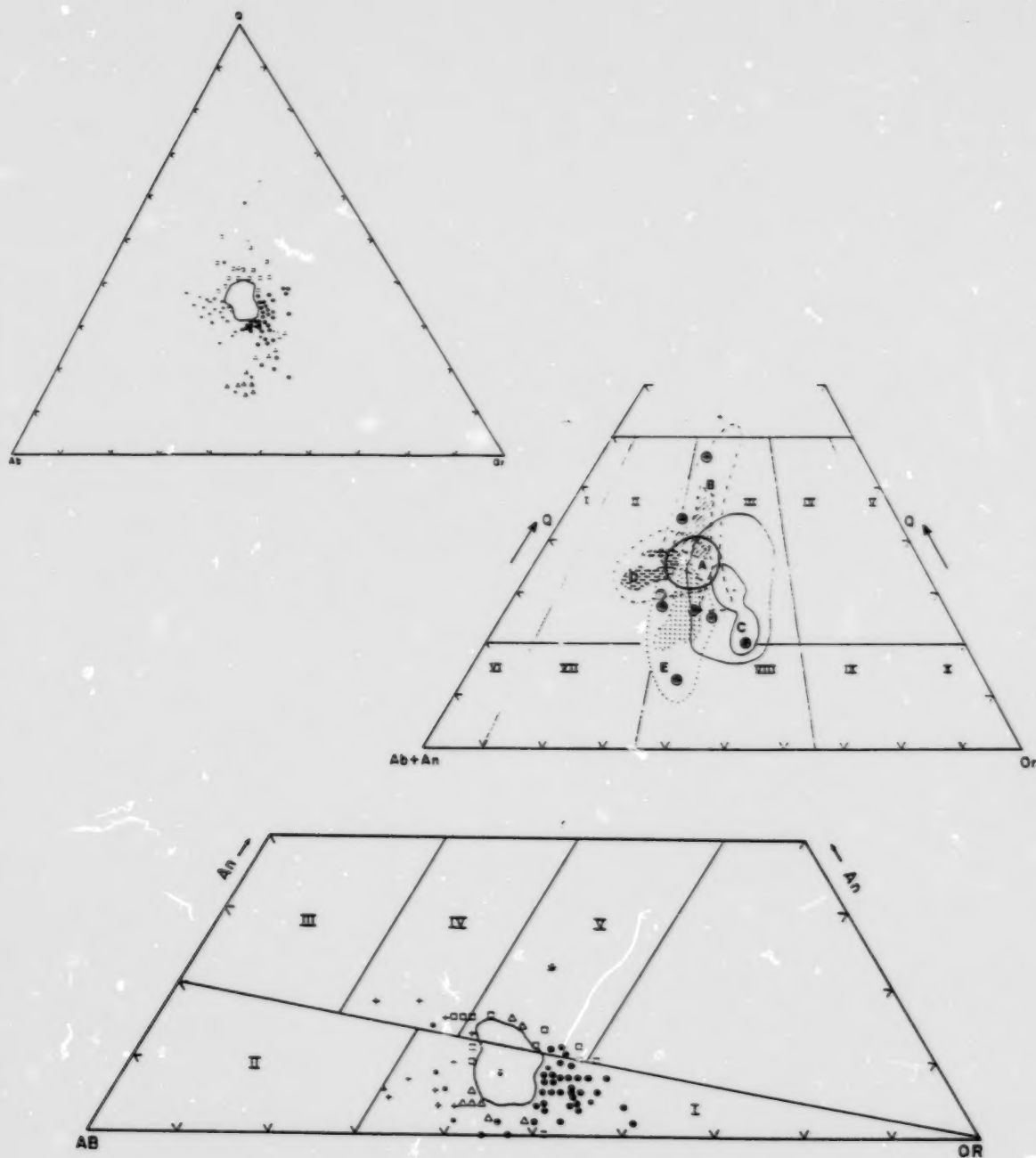


Figure 5-17. Normative Q - Ab - Or (A), Q - Ab + An - Or (B), and An - Ab - Or (C) diagrams showing the composition of the granodiorites of the Millet Brook area. In Figures (A) and (C) area enclosed with stippled pattern represent field of unaltered granodiorite. $\square$ = silicified granodiorites, + = albitized granodiorites, $\circ$ = K-feldspathized granodiorites, Δ = calcic granodiorites and star with circle represents distribution of mineralized samples. In Figure (B) patterned areas designated as a, b, c, d and e represent fields of unaltered granodiorite, silicified granodiorite, potassic granodiorite, albitized granodiorite and calcareous granodiorites respectively. Star with circle represents distribution of mineralized samples.

Table 5-3. Summary of statistical data for the 179 granitoid rocks, Millet Brook area. Mean values for major element oxides (per cent by weight); for trace element (ppm); $\bar{x}$ = arithmetic mean, S^2 = variance; $Cv\%$ = coefficient of variation $S/\bar{x} \times 100$; Ku = kurtosis; Sk = skewness; 95% CI = 95% confidence limit for mean value ($S^2 = t_{95}/\sqrt{n} \times Cv$, t_{95} Student's t value at 95% confidence level.

ELEMENTS	$\bar{x}$	S^2	$Cv\%$	δ	Ku	Sk	Range	95% CI
SiO_2	69.55	6.21	3.58	2.49	2.28	0.81	61.12 - 80.38	69.18 - 69.91
TiO_2	0.61	0.05	35.69	0.22	- .07	- .54	0.07 - 1.17	0.58 - 0.64
Al_2O_3	14.63	1.13	7.27	1.06	5.07	- .93	8.74 - 17.43	14.48 - 14.79
Fe_2O_3	0.52	0.08	54.58	0.29	1.42	1.15	0.02 - 1.56	0.48 - 0.57
FeO	3.16	0.72	26.98	0.85	1.58	-1.10	0.27 - 5.14	3.03 - 3.28
MnO	0.06	0.000	23.98	0.01	0.43	-0.46	0.02 - 0.10	0.06 - 0.07
MgO	1.06	0.08	27.40	0.29	3.66	-0.28	0.04 - 2.45	1.02 - 1.11
CaO	1.51	0.25	33.29	0.54	-0.68	-0.22	0.31 - 2.62	1.44 - 1.59
K_2O	4.31	0.57	17.49	0.75	0.77	0.38	2.16 - 6.82	4.20 - 4.42
Na_2O	3.21	0.10	9.98	0.32	2.22	-0.12	1.77 - 4.13	3.17 - 3.26
H_2O^+	1.04	0.07	25.52	0.27	2.16	0.95	0.42 - 2.21	1.01 - 1.08
H_2O^-	0.10	0.01	103.23	0.11	12.65	2.63	< 0.01 - 0.78	0.08 - 0.12
P_2O_5	0.22	0.004	28.35	0.06	31.05	3.41	0.01 - 0.74	0.21 - 0.23
CO_2	0.12	0.02	115.30	0.14	26.86	4.24	0.02 - 1.29	0.10 - 0.14
Ba	618	68780	42.45	262.0	3.72	1.43	105 - 1800	579 - 656
Cu	26	2895	206.16	54	27.60	4.95	< 1.0 - 410	18 - 34
F	492	12061	22.30	110	1.70	-0.42	100 - 890	476 - 508
Mo	2	0.86	42.67	0.93	1.87	1.06	0.5 - 5.7	2 - 2.3
Pb	31	777	90.44	28	69.84	7.38	10 - 320	26 - 35
Rb	167	516	13.58	22	-0.33	-0.09	103 - 221	163 - 170
Sn	5	3	34.30	1.74	42.0	5.54	3 - 20	4 - 5
Sr	130	787	21.57	28	-0.19	-0.23	52 - 195	125 - 134
Zn	138	9881	71.60	99	11.54	2.93	34 - 720	124 - 153
U	16	3370	361.87	58	143.94	11.48	1.4 - 748	8 - 24
B	23	452	89.61	21	14.77	3.19	3 - 170	20 - 26
Be	4	4	54.43	2	47.4	5.16	0.8 - 23	3 - 4
Bi	0.05	0.03	323	0.2	118	10.27	0 - 2.0	0.02 - 0.07
Li	83	499	27	22	1.46	-0.69	16 - 139	80 - 86
Co	4	14	110	3.8	-1.41	0.44	0 - 12	3 - 4
Ni	6	40	110	6.3	-1.35	6.34	<1 - 21	5 - 7
Th	10	11	34	3.3	11.37	0.25	8 - 31	9 - 10

In other words, if a major part of the compositional variability is due to the separation or redistribution of characteristic element assemblages, the observed compositional structure places constraints on the nature of the compositional variations within these elemental associations. Also, the results based on quantitative methods have merit in that they are objective and repeatable.

Using the SPSS 'Discriminant' program a single function was calculated to distinguish between the unaltered and the altered granodiorite, and the function is:

$$DF = 0.365 \log \text{SiO}_2 + 0.554 \log \text{TiO}_2 - .309 \log \text{Fe}_2\text{O}_3(\text{T}) - 0.549 \log \text{MnO} - 1.622 \log \text{CaO} - 0.510 \log \text{Cu} + 0.264 \log \text{F} - 0.991 \log \text{Rb} + 0.625 \log \text{Sn} + 0.952 \log \text{U} + 0.489 \log \text{B} - 0.570 \log \text{Be} - 0.408 \log \text{Ni}.$$

The eigenvalue for this calculated function is 2.438, with a canonical correlation of 0.842, a Wilks Lambda of 0.291 and Chi-squared of 56.805. Thus, judging from the eigenvalue and its canonical correlation it is obvious that the discriminant function provided an important measure to separate the two groups of rocks, i.e. unaltered granodiorite and the altered granodiorites. However, as discussed by Chatterjee and Strong (1984b), while there is no overlap between the two groups, one of the main mathematical criteria for reliable discrimination is not satisfied, namely that within-group variance is greater than that between the two groups. The separation of various types of altered rocks and the unaltered granodiorite was achieved through use of step-wise discriminant analysis. In this model, one variable is included to calculate a discriminant function in cumulative steps, this variable being the one that results in the most significant L value after adjusting for the variables already included in the model. At each step the computer output consists of the overall L value, the partial L for each variable already in the model (L to remove), the partial L for each variable still a candidate for inclusion, the discriminant function coefficient and the classification table. The four alteration types and the unaltered granodiorite were thus used to calculate the four discriminant functions (Table 5-4). Because the first two functions account for more than 60% of the total variance in the data set, the discriminant scores were plotted as a binary plot (Fig. 5-18). It can be seen from the above figure that each group has completely separate centroids and there is overlap of only two samples which may have been petrographically misclassified. It is important to note that this consistency of classification cannot be attained by the straight forward plotting of the individual element concentrations.

R-mode analysis was used to test and further develop the chemical processes responsible for the variability in the geochemical data (Table 5-2) from the Millet Brook area. In simple terms a factor (F) is a weighted linear combination of the original variables. In other words, it combines the basic variables into one or more composite variables which summarize the essential

Table 5-4

ELEMENT	FUNC. 1	FUNC. 2	FUNC. 3	FUNC. 4
Log SiO ₂	-0.703	-0.117	0.391	0.917
Log Al ₂ O ₃	-0.388	-0.651	0.496	0.360
Log MnO	0.124	0.233	0.226	0.679
Log CaO	-0.107	-0.924	-0.321	0.159
Log K ₂ O	-0.434	-0.195	-0.054	0.176
Log Na ₂ O	0.262	-0.060	-0.212	0.678
Log P ₂ O ₅	0.405	0.051	-0.077	-0.214
Log CO ₂	0.406	0.964	-0.355	-0.337
Log Ba	0.490	0.936	-0.714	0.579
Log Mo	0.188	0.106	-0.346	0.525
Log Pb	0.154	1.301	-0.350	0.314
Log Rb	0.352	-0.080	-0.002	-0.051
Log Sn	-0.430	0.079	0.239	-0.034
Log Zn	-0.389	-1.368	-0.113	-0.061
Log U	-0.298	0.250	0.180q	0.623
Log Be	0.109	-0.089	0.112	-0.676
Log Ni	-0.304	-0.282	0.385	-0.078
Log Th	-0.012	-0.445	-0.087	0.509
EIGENVALUE	1.311	0.809	0.697	0.614
% VARIANCE	38.22	23.57	20.31	17.91
CUMULATIVE %	38.22	61.79	82.09	100.00
CANONICAL CORRELATION	0.753	0.669	0.641	0.617
WILKS LAMBDA	0.088	0.202	0.365	0.619
CHI-SQUARED	134.04	120.80	75.065	35.155

information contained in the data set. Thus, factor analysis is helpful in the task of reducing the number of measurements needed to describe a phenomenon and because this is related to geological aspects, the interpretation of the factors is largely non-statistical.

As shown in Table 5-5 and Figure 5-19, first factors account for more than 85 per cent of the total variance in the data set. Also, the eigenvalue-factor curve shown in Figure 5-19 would suggest that essentially three distinct processes are the cause for the variance and these are marked by three distinct slopes. Factor I has an eigenvalue of 8.3 (first distinct process), in factors II, III and IV the eigenvalues are 2.8, 2.2 and 1.25 respectively (second distinct process) and for the remaining factors the eigenvalue range from 0.75 to 1.21 (third distinct process).

The positive factor loadings pertaining to the first factor as shown in Table 5-5, are characterized by TiO₂, FeO, MnO, MgO, CaO, H₂O*, F, Sr, Li and Th. The elemental associations clearly suggest a biotite factor. Thus, the biotite constituents cause most of the chemical variability in the data section and as discussed previously the composition of biotite from the Millet Brook area show systematic variations and form chemically distinct groups in different rock types. From the factor loading, it is also clear that Li and Th are tied up in the biotite structure and this in turn results in a significant positive correlation between Mg and Li. The CaO and Sr are probably related to the plagioclase in the granitoid rock and the Al₂O₃ has a low positive loading because its role is common in both the unaltered and altered rock

Table 5-5. Numerical values for various factors determined for the granodiorites of Millet Brook area. The calculations are based on the basis of the geochemical data presented in Table 5-2. The factors were rotated after Kaiser normalization.

	FACTOR I	FACTOR II	FACTOR III	FACTOR IV	FACTOR V	FACTOR VI	FACTOR VII	FACTOR VIII
SiO ₂	-0.505	-0.658	0.318	-0.295	-0.191	0.057	-0.073	0.092
TiO ₂	.779	.195	-.371	.135	.038	.040	.144	-.024
Al ₂ O ₃	.162	.820	-.062	.388	.094	-.072	-.038	-.053
Fe ₂ O ₃	.085	-.007	-.268	-.143	.178	.530	-.139	.238
FeO	.930	.127	-.210	.077	.083	-.165	.112	.014
MnO	.842	.175	-.135	-.127	.069	-.002	.064	-.117
MgO	.910	.071	-.201	.102	.131	-.055	.113	.094
CaO	.589	.276	-.276	.188	-.258	-.214	.420	-.017
K ₂ O	-.395	.089	.136	.638	.119	-.243	-.370	-.304
Na ₂ O	.022	.827	-.022	-.121	.047	-.061	.033	.200
H ₂ O ⁺	.436	-.014	.153	-.004	.374	.343	.036	.200
H ₂ O ⁻	.016	.004	.026	-.030	.029	.013	-.040	.146
P ₂ O ₅	.056	.176	.041	.019	.055	.081	-.023	.320
CO ₂	.050	.152	-.147	.067	.147	+.476	.384	-.017
Ba	.285	.138	-.236	.784	.072	-.066	.174	-.097
Cu	-.072	-.005	.051	.038	.152	.578	.028	.115
F	.826	.066	-.027	.032	-.005	.036	.011	-.132
Mo	.146	-.104	.120	.002	.170	.118	.709	-.067
Pb	-.062	.159	-.177	.235	.668	.118	.161	.034
Rb	.268	-.001	.115	.031	.220	-.046	-.362	-.608
Sn	-.204	-.016	.562	-.104	.120	.107	.012	.073
Sr	.409	.336	-.290	.427	-.048	.031	.167	.007
Zn	.256	.061	-.022	-.098	.798	.196	.130	.110
U	-.274	-.108	.195	-.104	-.015	.732	.018	+.919
B	-.112	-.205	.373	-.131	.152	.465	.031	.503
Be	-.139	-.024	.614	-.097	-.212	.051	.009	-.038
Bi	-.268	-.146	.657	.063	-.096	-.091	-.130	.305
Li	.868	-.030	-.125	-.083	.130	-.107	-.181	-.039
Th	.835	-.040	-.044	.180	-.008	-.143	.089	.163

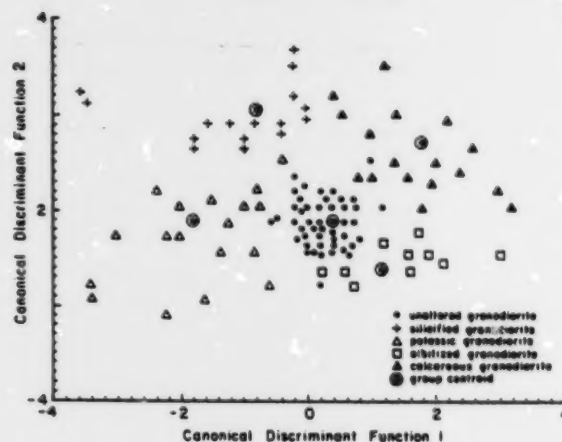


Figure 5-18. A plot of the discriminant scores using the first two function calculates using discriminant analysis. For numerical values of the variables see text.

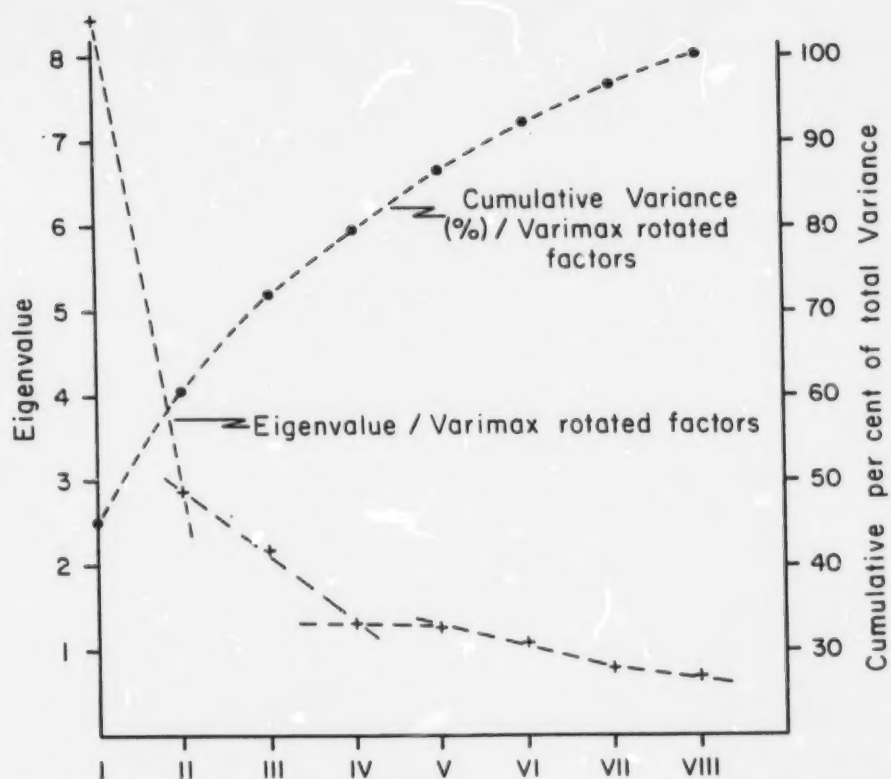


Figure 5-19. Factor - eigenvalue diagram for the granodiorites of the Millet Brook area. The cumulative variance (%) curve is also shown for reference.

groups. Thus, the chemical variability of the granodiorites appears to be caused by two minerals, i.e. biotite and plagioclase feldspar, by trace elements incorporation in the lattice framework of these minerals. The conclusion is entirely in agreement with previous studies, by Clarke and Muecke (1981) and Chatterjee and Strong (1982b), regarding the magmatic cocrystallization of biotite and plagioclase feldspar in the early magmatic history of the SMB.

Factors II, III and IV may be grouped together as the hydrothermal alteration stage. The factor loading clearly illustrates the role of albitization (Na_2O , CaO , Al_2O_3 , Sr), silicification (SiO_2 , Be, B_2O_3) and K-feldspathization (K_2O , Al_2O_3 , Ba, Pb, Sr). This interpretation is consistent with the petrographic and mineralogical observations discussed previously and also illustrated by discriminant analysis.

The remaining factors may be termed the hydrothermal mineralization stage where essentially the factor loadings are dominated by U, CO_2 , B, P_2O_5 , Cu-Pb-Zn. This is clearly the case when the mineralogical make-up of the uraniumiferous zones are considered, e.g. association of chalcopryrite with uranium phosphates/pitchblende, association of calcite and pitchblende etc., and also the overall elemental distribution and the covariance of B and Li is considered (Clarke and Chatterjee, 1985).

Thus, from the preceding discussions the following may be concluded:

(a) the chemical variability in the granodiorites of the Millet Brook is controlled by magmatic and postmagmatic processes (Clarke and Chatterjee, 1985) leading to hydrothermal U - Cu - B - $\text{CO}_2 \pm \text{P}_2\text{O}_5$ mineralization,

(b) the association of elements appears to be a sensitive indicator of the elements concentrated in the hydrothermally altered rocks, hence potentially useful as a pathfinder elements for mineral exploration, and

(c) though the mechanisms responsible for uranium transport remain speculative, some insight may be gained from the close association of the volatiles, e.g. CO_2 and B, and perhaps B is more important than the commonly documented role of CO_2 and F as a complexing agent (cf. Clarke and Chatterjee, 1985).

Rare Earth Elements and Whole Rock Oxygen Isotope Relationships

Variation in rare element concentrations is becoming increasingly recognized as a means of detecting and understanding the subtle effects of processes responsible for mineralization in deposit types ranging from volcanogenic massive sulphide (Campbell et al., 1982; Strong, 1984) to porphyry copper (Taylor and Fryer, 1982), greisen-type tin (Chatterjee and Strong, 1984a) and tungsten (Higgin, 1980) deposits and sediment-hosted uranium deposits (McLennan and Taylor, 1979) and vein-type uranium deposits in the granitoid rocks of the SMB (Chatterjee and Strong, 1985b).

The REE data from the Millet Brook uranium deposits is plotted in Figure 5-20. The unaltered rocks exhibit fairly consistent patterns like those of the granodiorite of the SMB anomalies (i.e. negative slope and negative Eu). The REEs range from 115 to 135 ppm, the U/Th ranges from 0.26 to 0.49 and δ^{180} ranges from 10.18 ‰ to 10.86 ‰ (Fig. 5-21).

The altered granodiorites have more gently sloping patterns than the unaltered, with an overall drop in the light REE (LREE), a slight decrease in the size of the Eu anomalies and an increase in the heavy REE (HREE), whereas the intermediate REE (Sm, Eu, Gd) remain more or less constant. This crossing of the patterns for altered and unaltered rocks is not seen in the suites studied by Muecke and Clarke (1981), and is interpreted to be the result of a more intense effect of different kinds of fluid transport and deposition as suggested by these authors. Given the predicted effects of different kinds of fluids as summarized by Taylor and Fryer (1982, 1983), it might be suggested that the drop in the LREE results from leaching effects of Cl-bearing fluids, and the HREE increase results from deposition by CO_2 or F-bearing fluids. The altered rocks also have a wider range of REE variation than do the unaltered, as would likewise be expected as a result of fluid activity. The U/Th ratio for the altered rocks ranges from five to ten and the whole rock δ^{180} values show an increase from 11.12 to 12.38 ‰, thus establishing a strong positive correlation between U/Th ratio and δ^{180} values of the rock and a negative correlation between ΣREE and δ^{180} .

The mineralized samples exhibit several types of changes in the REE patterns relative to the altered or unaltered granodiorites, which correspond to the two types of alteration indicated on Figure 5-20. The most striking change is the complete loss of europium from the mineralized samples from the silicified zone. This clearly results from destruction of feldspars, a feature also noted by Alderton et al. (1980) for alteration of granodiorites associated with mineralization in southwest England. It can be seen that there appears to be several stages to the alteration effects as the silicified zone of mineralization is approached. Firstly, from the unaltered sample to the albitized rock, there is a major drop in LREE and increase in HREE. Then, within the mineralized rocks, there is a progressive increase in the LREE back to normal levels and a slight drop in HREE. Next, within the two silicified samples, there is first a drop in La + Ce + HREE (A16-9) and then an increase in the total REE in A16-10.

In the desilicified (episyenitized) zone, the most uranium-enriched sample containing 1.1% U (A16-13; not shown in the diagram) has also lost all its europium, like those of the silicified zone, but has been relatively enriched in the HREE, as would be expected from the fact that it also contains 10.68% CO_2 .

The mineralized granitoid rocks, as expected, show a progressive increase in the U/Th ratio which reaches a high of 580. However, they

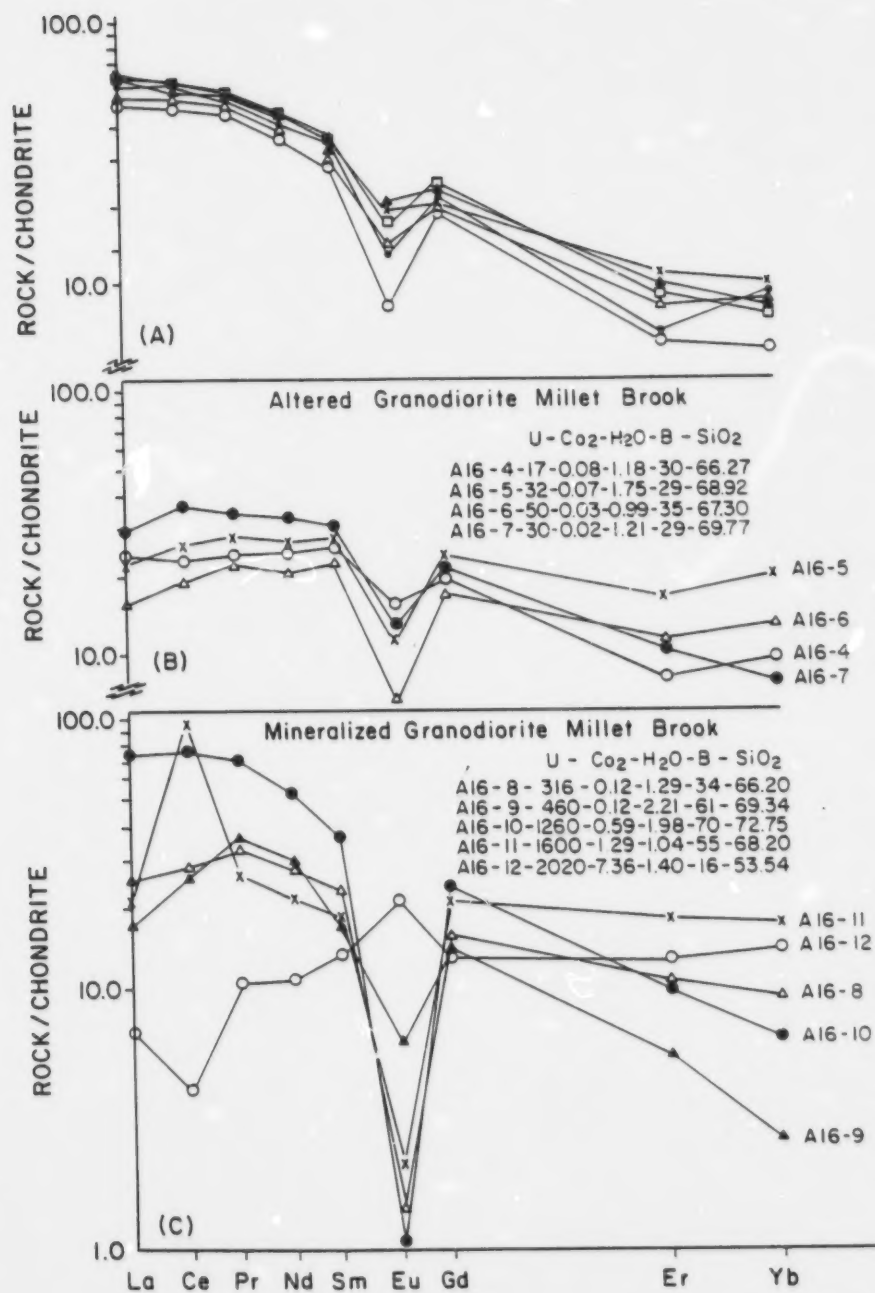


Figure 5-20. Chondrite normalized REE patterns for the unaltered, altered and mineralized granodiorites of the Millet Brook area.

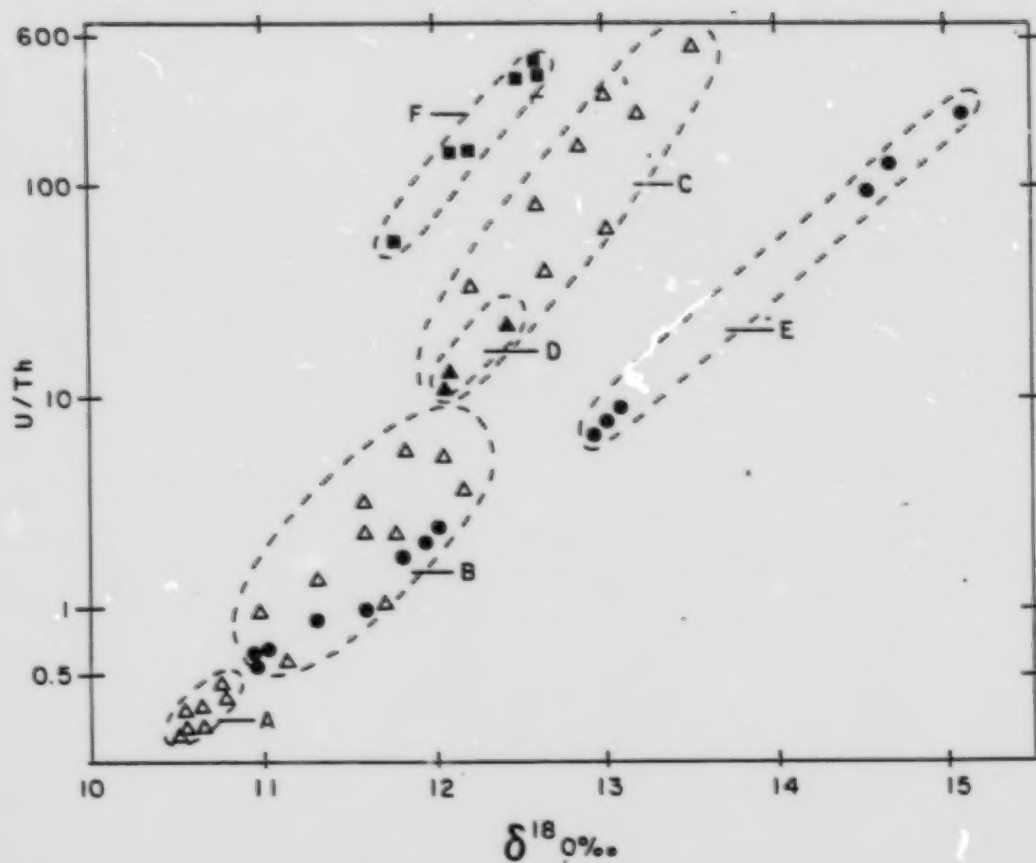


Figure 5-21. U/Th vs whole rock $\delta^{18}\text{O}$ ‰ variation diagram showing the distribution of unaltered rocks (A) altered rock (B) and mineralized rocks (C). Field (D) defined by solid triangles represent uraniferous leucogranite dykes.

also show an increase in whole rock δ^{180} values which range from 1.23 to 15.3 ‰ (Fig. 5-21), representing a positive shift of approximately 4.44 ‰ from the unaltered rocks with background abundance of U, Th, K and REE. Other uranium deposits within the SMB show similar positive correlation between U/Th ratio of the rock and whole rock δ^{180} values. However, samples from each deposit define a slightly different slope. These results are consistent with the pervasive access of hydrothermal fluids to the wall rocks and probably represent mixing of metamorphic and internally generated magmatic fluids.

The data and the discussion presented in the preceding pages demonstrate that the source for the uranium in the deposits was not the host rocks and the uranium in the veins could not have been deposited by mobilizing uranium from the wall rocks. Instead uranium has been added to the wall rocks. We will present data demonstrating that the late stage uraniferous leucogranites have lost approximately 70% of their original uranium and within the deposits these rocks are represented in the form of uraniferous 'feeder dykes'.

The late stage uraniferous leucogranite dykes are characterized by strong depletion of the LREE, and a total lack of Eu and the HREE. We do not feel that this can be explained by any of the fluid alteration effects discussed above because there is no petrographic evidence that the dyke has been significantly altered. We suggest, rather, that this pattern reflects an extreme example of magma depletion by fluid loss which became increasingly important in the late stages of cooling of the Batholith, a stage at which the magma may have been already depleted of the HREE by the loss of such fluids, especially CO_2 - or F-rich varieties. We note that such an interpretation has also been made by Alderton et al. (1980) for REE-depleted late stage fluorites in Cornwall, and by Strong et al. (1984) for progressively younger stages of zoned fluorite veins at St. Lawrence, Newfoundland. The U/Th ratio for the uraniferous dykes is 15 ± 2 and the δ^{180} value is approximately 12 ‰ (field D in Fig. 5-21).

Source of Uranium

The geological, petrological, mineralogical and geochemical arguments presented in this paper, clearly suggest that the hydrothermal alteration and mineralization has a superimposed nature, i.e. the mineralizing fluids were later emplaced in the crystalline host rocks. However, to access the problem of source or sources for the metals, current theories of vein type uranium mineralization invoke either the conventional magmatic-hydrothermal model or variant of the above involving addition or interaction of ground water or formation waters during alteration and mineralization. In the former, the sources of metals and the bulk of hydrothermal fluids are considered to have an igneous origin, being derived from portions of granitic rocks deeper than, or adjacent to, the mineralized zone. In the latter, varying amounts of meteoric-type waters from external sources participate and

recognition of this is largely based on oxygen and hydrogen isotopic data. Chatterjee and Strong (1985a,b) using whole rock oxygen isotope data have suggested that perhaps the metamorphic waters are equally important. In all of these models, emphasis is placed on the nature of the fluids and many of these studies do not present clear evidence regarding the depletion of the element(s) from the source rocks(s), though the evidence for the concentration of the element(s) in the deposit is obvious. To fill this gap we have investigated the vertical distribution of radioelements in the drill core and in combination with the fission track studies we were able to demonstrate the loss of uranium from the biotite leucomonzogranite of the New Ross area. The drillholes are located at Turner tin prospect (T7) and at Lake Lewis (AV-1). The analytical data is given in Table 5-6 and plotted in Figure 5-22 and 5-23. The uranium distribution in rocks is illustrated in Figure 5-24. Some of the essential characteristics of the radioelement profiles are as follows (for detail discussion, see Chatterjee and Strong, 1985b).

(a) The leucomonzogranite are characterized by uniform grain size and the intensity of the fractures increases with decreasing depths. The fracture zones are characterized by hematitic alteration of the K-feldspars and sericitization is generally restricted adjacent to the fracture zones. The geophysical logs (available only for AV-1) correlate elegantly with the physical and chemical characteristics of the granite rocks.

(b) The distribution of uranium and thorium clearly shows that with increasing depth uranium increases whereas the Th levels remain more or less constant. In both the drillholes, the change is clearly marked below the ~100 metre depth and it must be emphasized that this change is not related to the change in rock type. The profiles also show that the fracture zones are enriched in uranium, thus clearly establishing the concentration of uranium in fractures.

(c) The Th/U ratio is markedly different at a depth of approximately 100 metres. It is clear that the ratio increases with decreasing depth probably due to the loss of uranium and also due to slightly different behaviour of uranium and thorium in the hydrothermal fluids (Chatterjee and Muecke, 1982, Chatterjee and Strong, 1984a). If the Th/U ratio reflects preferential loss of uranium then it must be concluded that the leucomonzogranite has lost approximately 70 per cent of its original uranium from the top 100 metres (at least in the present case).

(d) The radioelement profiles do not present any physical or chemical evidences for loss or gain of uranium, they simply show the distribution of abundances. To demonstrate the loss of uranium it is essential to document its nature and distribution in the rock. This is achieved by the fission track method and is illustrated in Figure 5-24. In the following discussion textural features are described which have bearing on the uranium migration, retention and loss.

In addition to common rock-forming minerals the leucogranite has a large variety of accessory

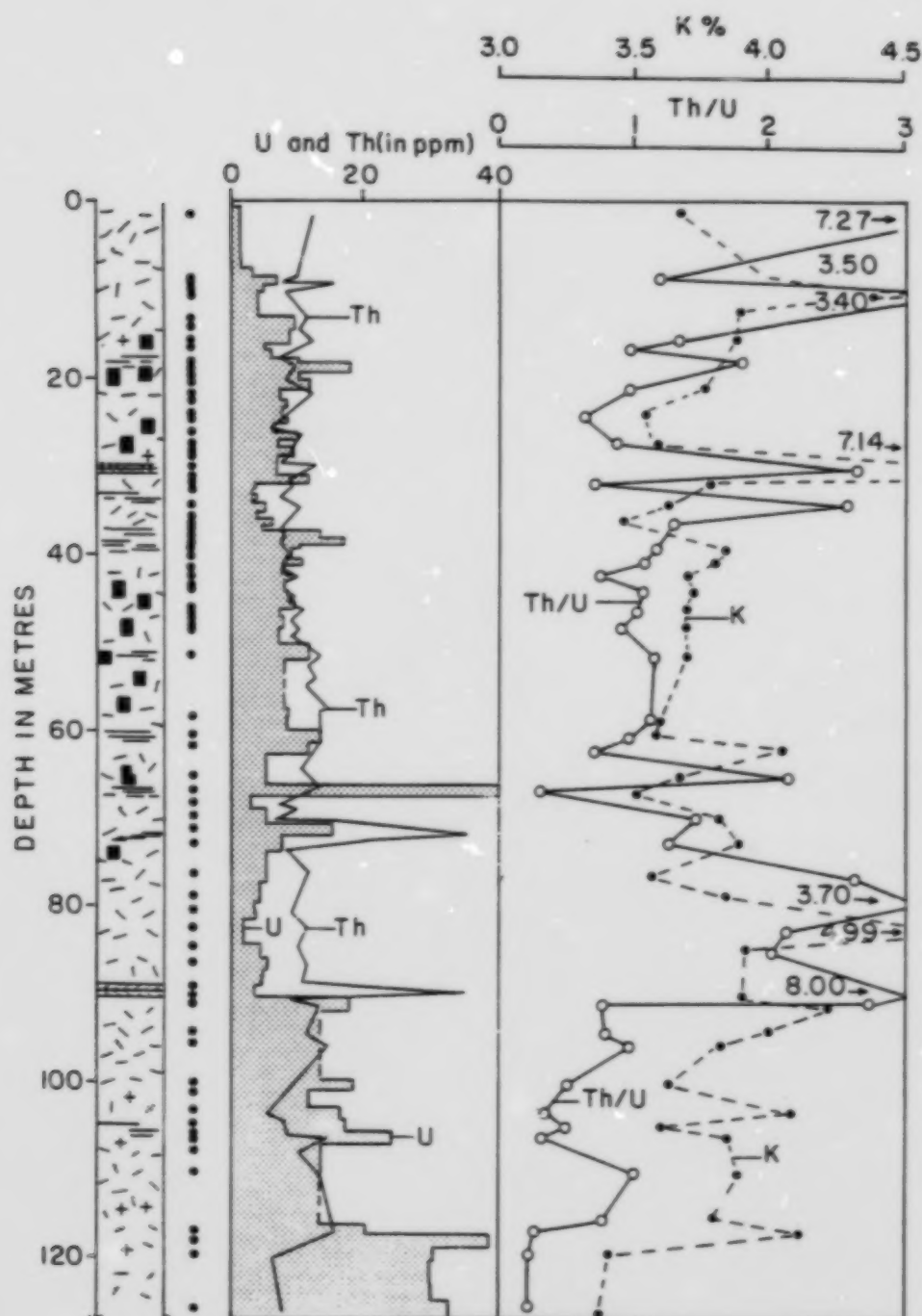


Figure 5-22. Vertical distribution of U and the Th in a drillhole (DDH + T7) from New Ross area (for explanation see text). Dots represent sample location and closely spaced line in the first column designate fractures. The solid squares represent intense hematite alteration of feldspars.

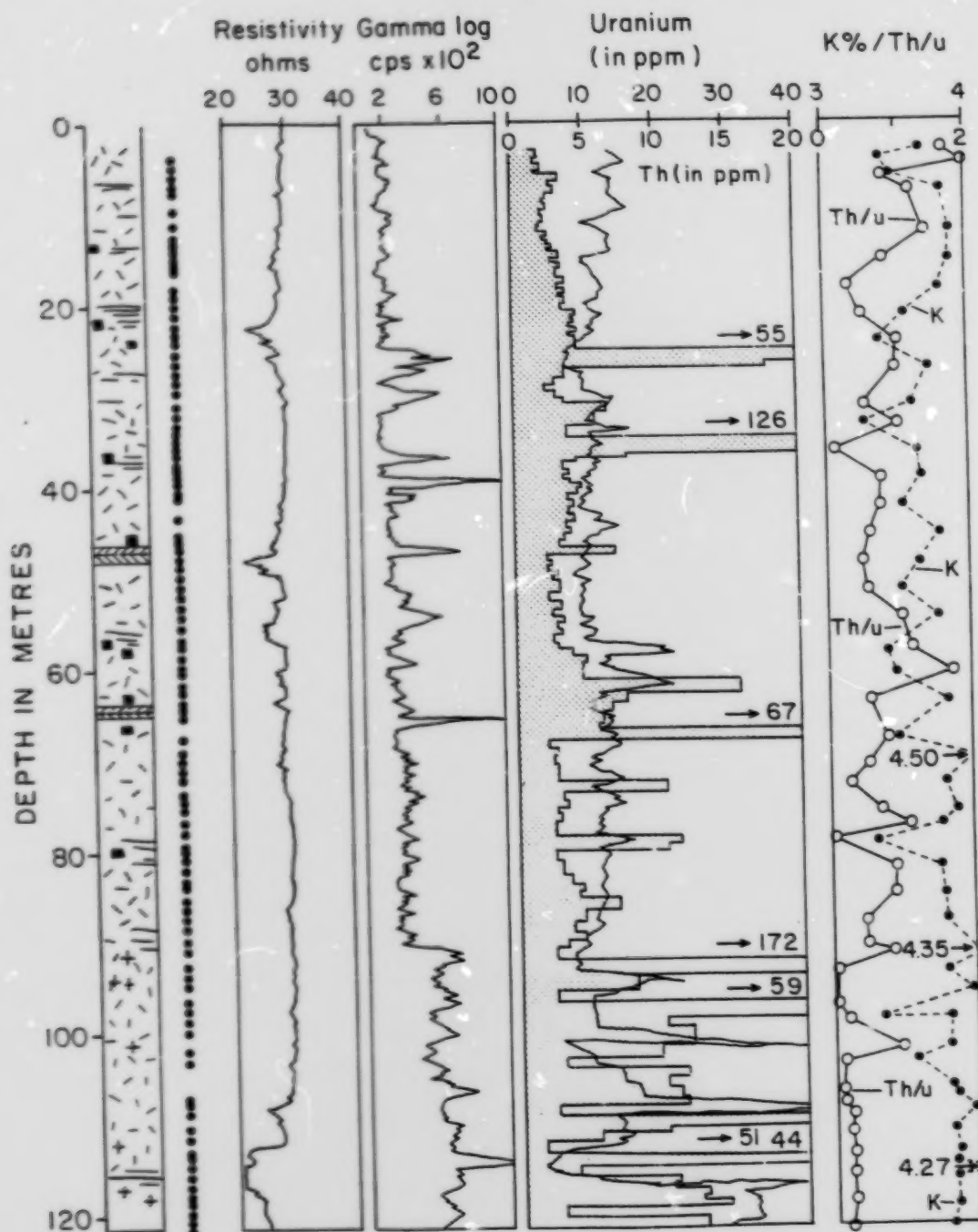


Figure 5-23. Vertical distribution of radioelements from Lake Lewis leucomonzogranite. The resistivity and gamma logs are given for reference. Patterned zones in the lithologic units designate shear zones with intense alteration.

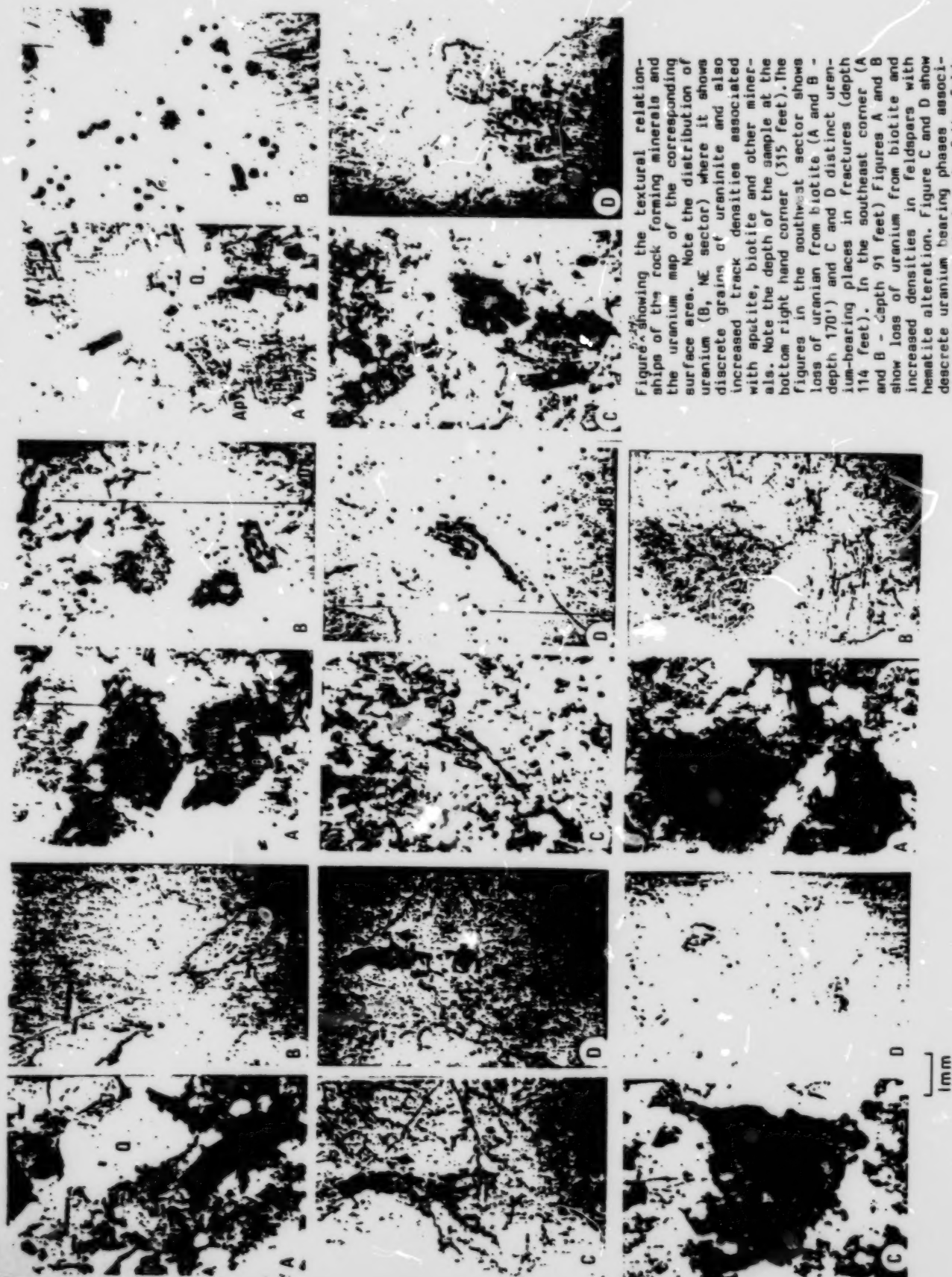


Figure 6. showing the textural relationships of the rock forming minerals and the uranium map of the corresponding surface area. Note the distribution of uranium (B, NE sector) where it shows discrete grains of uraninite and also increased track densities associated with apatite, biotite and other minerals. Note the depth of the sample at the bottom right hand corner (315 feet). The figures in the southwest sector show loss of uranium from biotite (A and B - depth 170') and C and D distinct uranium-bearing places in fractures (depth 114 feet). In the southeast corner (A and B - depth 91 feet) Figures A and B show loss of uranium from biotite and increased densities in feldspars with hematite alteration. Figure C and D show discrete uranium bearing phases associated with the marginal part of the feldspar grains (depth 31 feet).

DBI No. Depth in Feet	U	Th	K	Th/U	DBI No. Depth in Feet	U	Th	K	Th/U	DBI No. Depth in Feet	U	Th	K	Th/U
T7-5	1.55	12	3.68	7.27	T7-193	8.10	9	3.56	1.11	AVI-44	6.4	5	3.66	0.78
T7-20	2.95	10	3.98	7.50	T7-200	13.40	13	3.56	0.97	AVI-50	5.0	5	3.90	0.83
T7-21	6.30	7	4.39	1.11	T7-204	10.75	7	4.06	0.65	AVI-51	6.4	5	3.52	0.71
T7-31	6.40	15	3.81	3.60	T7-215	5.10	11	3.67	2.15	AVI-56	6.4	4	3.68	0.63
T7-33	4.40	6	3.92	1.73	T7-220	41.20	12	3.50	0.29	AVI-61	16.4	6	3.82	0.37
T7-44	3.65	11	3.98	3.01	T7-225	2.95	7	3.61	1.77	AVI-62	5.0	5	3.66	0.83
T7-46	8.95	10	3.82	1.34	T7-230	4.70	7	3.81	1.68	AVI-65	5.8	4	3.16	0.69
T7-51	8.90	12	3.95	1.55	T7-235	14.30	37	3.32	1.22	AVI-66	9.4	5	3.52	0.53
T7-53	6.65	10	4.13	2.12	T7-240	6.55	8	3.87	2.60	AVI-67	5.4	5	3.56	0.82
T7-54	5.70	8	4.00	0.95	T7-252	4.60	12	3.35	2.65	AVI-71 ⁶⁴	6.1	5	4.19	0.74
T7-55	5.70	10	4.00	2.12						AVI-73	5.4	5	3.56	0.82
T7-56	8.60	8	3.59	1.81						AVI-75	5.4	5	3.52	0.94
T7-59	8.60	9	3.76	1.81						AVI-77	6.6	5	3.66	0.87
T7-60	4.95	10	3.78	0.63						AVI-77 ⁶⁴	5.3	5	3.59	0.94
T7-62	15.55	7	3.87	1.01						AVI-80	4.0	4	3.27	1.00
T7-64	6.90	10	3.87	1.01						AVI-83	8.3	4	3.65	0.48
T7-65	9.58	9	3.84	0.93	T7-260	4.05	11	3.82	2.65	AVI-85	3.8	7	3.67	1.84
T7-69	11.07	11	3.98	0.99	T7-265	3.85	11	3.86	2.85	AVI-86	55.9	5	3.66	0.09
T7-71	5.60	10	3.75	1.78	T7-271	2.70	10	4.89	3.70	AVI-87	36.4	3	3.34	0.08
T7-72	9.65	7	3.53	0.74	T7-279	4.40	9	3.89	2.04	AVI-89	5.0	5	3.75	1.00
T7-74	1.35	9	4.06	5.66	T7-286	5.61	10	3.84	2.00	AVI-91	4.9	4	3.64	0.82
T7-75	8.80	6	3.79	0.68	T7-293	4.25	34	3.81	8.00	AVI-96	5.2	3	3.34	0.58
T7-80	9.25	5	3.53	0.60	T7-297	2.90	8	3.90	2.71	AVI-97	4.4	4	3.42	0.91
T7-81	5.20	5	3.68	0.96	T7-300	16.00	13	4.28	0.76	AVI-102	6.0	3	3.60	0.50
T7-85	6.10	4	3.80	0.65	T7-310	14.20	11	3.94	0.77	AVI-103	9.6	5	3.57	0.52
T7-90	6.00	8	3.54	0.88	T7-315	14.20	14	3.82	0.94	AVI-104	12.3	5	2.97	0.41
T7-91	7.05	7	3.46	0.99	T7-320	17.80	9	3.63	0.50	AVI-109	8.3	9	3.25	1.08
T7-92	3.80	9	3.54	2.36	T7-322	10.90	8	3.93	0.73	AVI-111	3.9	5	3.60	1.28
T7-93 ⁶⁴	3.95	1	3.59	2.78	T7-340	5.65	5	4.15	0.32	AVI-112	4.0	7	3.16	1.75
T7-94	4.40	10	3.43	2.27	T7-345	16.35	8	3.58	0.48	AVI-115	46.6	9	4.14	0.193
T7-95	3.65	11	3.55	3.18	T7-349	26.60	9	3.60	0.32	AVI-116	126.0	9	5.31	0.07
T7-100	3.75	10	7.14	2.66	T7-350	26.10	13	3.86	0.49	AVI-118	29.4	5	3.41	0.17
T7-104	5.70	13	3.76	2.28	T7-354	13.50	9	3.54	0.66	AVI-119	16.4	3	3.67	0.18
T7-105	12.70	9	3.78	0.70	T7-366	13.0	13	3.88	1.00	AVI-120	15.1	4	3.49	0.26
T7-107	3.20	7	3.72	2.18	T7-385	19.0	15	3.79	0.78	AVI-122	8.9	3	3.21	0.36
T7-114	3.60	9	3.99	2.66	T7-392	39.3	11	4.25	0.27	AVI-126	7.1	5	3.57	0.70
T7-117	3.72	8	3.73	2.15	T7-394	29.4	6	3.69	0.20	AVI-128	5.7	5	3.68	0.87
T7-118	8.00	8	3.71	1.00	T7-413	32.65	7	3.36	0.21	AVI-133	9.0	2	3.62	0.25
T7-120	7.75	10	3.44	1.29	AVI-13	4.0	7	3.73	1.75	AVI-135	5.2	2	3.27	0.38
T7-121	7.15	8	3.81	1.11	AVI-15	6.6	6	3.66	0.91	AVI-137	4.8	4	3.45	0.83
T7-122	2.75	10	2.61	3.63	AVI-25	5.9	7	3.85	1.19	AVI-144	5.1	7	3.47	1.37
T7-124	13.2	10	4.43	0.75	AVI-32	5.0	7	3.89	1.40	AVI-169	5.6	3	3.55	0.53
T7-125	16.90	9	2.03	0.53	AVI-37	5.0	7	3.89	1.40	AVI-150	4.5	3	3.80	0.66
T7-129	9.75	11	3.86	1.12	AVI-41	5.0	9	3.77	0.71	AVI-152	9.0	4	3.78	0.64
T7-133	8.15	6	3.95	1.10	AVI-42	5.3	5	3.72	0.71	AVI-153	5.4	4	3.73	0.74
T7-137	7.65	7	3.83	0.91	AVI-43	5.2	7	3.57	1.25	AVI-155	5.8	3	3.47	0.51
T7-140	10.55	8	3.58	0.75	AVI-46	5.6	7	3.36	1.25	AVI-158	9.3	3	3.65	0.32
T7-142	8.98	6	3.86	0.99	AVI-295	5.5	5	3.52	0.76	AVI-159	16.9	3	3.55	0.17
T7-159	6.6	3	3.73	0.46	AVI-299	5.2	4	4.11	0.90	AVI-160	25.1	21	4.1	0.86
AVI-160	5.6	2	3.52	0.45	AVI-300	5.2	4	4.35	0.83	AVI-161	28.0	21	4.7	1.53
AVI-161 ⁶⁴	5.7	3	3.18	0.52						AVI-162	25.2	23	3.8	0.91

Table 5-6: Distribution of U, Th, K, and Th/U ratio in the biotite bearing leucomonzogranite of the New Ross area. The sample number reflects the diamond-drillhole and the depth in feet.

Table 5-6: Continued.

Sample No.	Depth (m)	Grain Size (mm)	Th/U	Th/Pb
AVT-154	5.7	2	2.36	0.30
AVT-155	5.6	4	2.36	0.71
AVT-157	5.0	2	2.57	0.
AVT-158	4.0	3	2.09	0.42
AVT-172	6.3	4	2.55	0.47
AVT-172 ¹⁰⁰	7.1	4	2.34	0.56
AVT-172	7.1	2	2.44	0.42
AVT-172 ¹⁰⁰	6.2	4	4.17	1.28
AVT-174	4.2	4	2.50	2.14
AVT-174	4.3	5	2.75	1.18
AVT-184	7.4	4	2.75	0.81
AVT-188	4.4	3	2.40	0.55
AVT-189	8.4	10	2.39	1.10
AVT-160 ¹⁰⁰	4.4	11	2.32	2.20
AVT-184	4.0	7	2.26	1.01
AVT-187	4.4	4	2.40	1.91
AVT-204	4.0	4	2.43	0.47
AVT-205	22.0	4	2.74	0.14
AVT-209	7.1	5	2.22	0.20
AVT-220	7.2	7	2.55	0.91
AVT-210	17.0	11	2.45	0.41
AVT-212	12.7	7	2.90	0.51
AVT-221	9.0	7	2.44	0.78
AVT-224 ¹⁰⁰	47.4	4	2.40	0.08
AVT-227 ¹⁰⁰	5.5	4	2.46	1.03
AVT-228	6.9	3	4.30	0.43
AVT-234	4.3	7	2.22	0.86
AVT-234	6.1	5	2.44	0.91
AVT-240	21.3	5	2.40	0.23
AVT-242	6.2	4	2.71	0.94
AVT-242	0.6	4	2.35	0.93
AVT-244	2.4	4	2.70	0.48
AVT-251 ¹⁰⁰	5.3	9	2.90	0.72
AVT-254	5.2	6	2.74	1.15
AVT-254	22.5	1	2.23	0.04
AVT-252	21.4	6	2.49	0.14
AVT-253	5.3	3	4.35	0.54
AVT-254	4.6	5	2.48	1.09
AVT-254	4.2	7	2.73	1.12
AVT-270	4.5	6	2.75	0.92
AVT-272	1.4	3	2.59	0.51
AVT-274	3.4	6	2.50	2.48
AVT-291	6.2	2	2.80	0.47
AVT-292	10.1	3	2.51	0.21
AVT-294	4.2	10	2.40	0.47
AVT-299	5.5	3	2.92	0.46
AVT-301	7.1	4	2.57	0.56
AVT-303	4.0	4	2.34	0.93
AVT-307	172.0	9	2.75	0.04
AVT-313	15.5	8	2.80	0.51
AVT-314	4.0	11	2.75	1.37
AVT-315	4.3	4	2.73	0.95
AVT-320	54.7	3	4.04	0.09
AVT-322	59.6	4	2.78	0.13
AVT-327	24.9	4	2.32	0.34
AVT-325	19.0	4	2.70	0.21
AVT-337	4.0	4	2.44	1.09
AVT-333	22.8	4	2.46	0.17
AVT-354	5.3	4	4.01	1.13
AVT-354	24.4	7	2.51	0.24
AVT-342	31.0	7	4.10	0.13
AVT-343	12.4	4	4.35	0.47
AVT-374	14.7	5	2.46	0.34
AVT-380	24	3	2.41	0.21
AVT-383	4.1	5	4.27	1.21
AVT-384	2.4	5	2.52	1.74
AVT-385	11.3	8	2.95	0.44
AVT-387	11.9	4	2.55	0.50
AVT-388	44.5	7	2.74	0.15
AVT-390	23.4	6	2.40	0.25
AVT-391	24.1	7	2.94	0.11
AVT-393	39.7	6	2.46	0.17
AVT-397	4.1	7	2.42	1.14
AVT-398	24.5	5	2.74	0.20

minerals e.g. apatite, monazite, uraninite, coffinite, fluorite, thorite, rutile, ilmenite and sulphide minerals. The fission track 'mapping' has shown that there are four modes of microscopic occurrence of uranium: (a) in accessory minerals, (b) along grain boundary, (c) in microfractures and (d) in biotites, feldspars and quartz.

On the basis of abundance and distribution of accessory minerals and the fission track they produce, it is estimated that approximately 25 per cent of the uranium in the leucogranite is held in the accessory minerals. Some uranium also occurs along the outer portion of apatite and fluorite crystals. However, more significant in abundance and distribution, is uranium along the boundaries of biotite crystals. This mode of

occurrence is not uniform. Biotite crystals below the 100 metre depth are densely surrounded by fission tracks. Above this depth, they show progressive change from decreasing intensity to total lack of fission tracks. Often, this biotite shows chloritization and the track density is controlled by crystallographic orientation or along cleavage planes. It is evident the biotite has lost its uranium along crystal boundaries during hydrothermal alteration. Uranium in microfractures accounts for high uranium content of many samples. Generally the fractures contain secondary minerals together with autenite, metatorbernite and fluorite, indicating precipitation of uranium. Widely scattered fission tracks from quartz and feldspar indicate that some uranium occurs as atomic substitution in these minerals.

The fission-track uranium maps of the thin sections show some areas of increased track densities that do not coincide with identifiable, microscopic uranium-bearing minerals. Instead, they coincide with grain boundaries, cracks or alteration zones, and the actual location of the uranium could be in crystallographic sites of submicroscopic uranium-bearing minerals or in fluid inclusions, or the uranium could occur as ions or molecules sorbed on to a surface.

The movement of uranium in solution through the granitoids depends on a physically, as well as chemically, unobstructed pathway. The increased density of fission tracks, associated with fractures and cracks that contain secondary and alteration minerals, suggests that the cracks provided pathways for uranium-bearing fluids as well as access to the primary and secondary uranium sites by oxidizing and complexing solutions.

The most extensively documented case of uranium loss at different stages of evolution is for the 2.9 b.y old granitoid of the Granite Mountain, Wyoming. Judging from the fission-track uranium map for the Granite Mountain granitoid presented by Stuckless et al. (1977), the uranium phases have identical distribution to the presently studied leucogranites of the New Ross area, suggesting similar mechanism for their origin.

CONCLUSIONS

The U - Cu - B - P₂O₅ mineralization at Millet Brook occurs in alteration zones within the least differentiated biotite granodiorite of the SMB. The process of alteration and mineralization resulted in numerous chemical effects, and the association of elements and their patterns of variation and sensitive indicators of these processes and of the presence of mineralization.

The mineralizing fluids probably emanated from the uraniferous leucogranite uranium was transported in the solutions as carbonate complexes and its deposition resulted from CO₂ exsolution from this fluid and its reaction with the host rocks, in a manner analogous to that suggested by Cuney (1978) and Leroy (1978). The hydrous silicate + carbonate + phosphate + borosilicate + sulphide mineral assemblage produced from the mineralizing fluids implies that the fluid was dominantly H₂O-CO₂-B with some phosphorus and sulphur. The leucogranite displays evidence that suggests microscale leaching, migration and deposition, and transport of the uranium by way of fractures and occurred microcracks.

CHAPTER 6

PORT MOUTON PLUTON

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Woodend, S. L., Port Mouton Pluton; in Guide to the Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

INTRODUCTION

The Port Mouton granitoid Pluton (Fig. 6-1) intruded the Goldenville Formation of the Meguma Group in southwestern Nova Scotia. The country rocks consist predominantly of greywacke, minor argillite, slate and mica schist units and, in this part of the Province, the regional metamorphic grade is lower amphibolite facies. The dimensions of the Pluton are approximately 20 km (northeast-southwest) by 9 km (northwest-southeast). Two small plugs are reported to occur west-northwest of the main Port Mouton Pluton and are assumed to be associated with the Pluton.

The Port Mouton Pluton appears to be a very complex intrusion. At least eight distinct units with crosscutting relationships have been mapped. The correlation between the units from the western section of the Pluton (including Port Joli Bay, St. Catherine's River) and the eastern section of the Pluton (Western Channel, Port Mouton Island coastline) is proving to be difficult at this preliminary phase of investigation. Feldspar staining has not been completed for many samples, making correlation between the units that much more difficult. The term 'granitoid' will be used where the exact nomenclature of the unit is currently unknown. Fine grained granitoids refer to grain sizes of between 0.5 - 1.0 mm, medium grained to sizes between 1 and 4 mm and coarse grained of grain sizes greater than 4 mm. The term leucocratic refers to a mafic mineral content of less than 5%, mesocratic refers to the mafic mineral content of between 10% and 20%, mafic refers to the mafic content of greater than 20%.

Radiometric ages for the Pluton vary. An age of 349 ± 15 Ma was obtained by Cormier and Smith (1973) using the Rb/Sr whole rock method. Reynolds (1980) obtained an age of 300 ± 8 Ma using the $^{40}\text{Ar}/^{39}\text{Ar}$ method. Elias (1985, personal communication) has recently measured an age of approximately 340 Ma on a muscovite from a late pegmatite in the Pluton. Since the Port Mouton Pluton appears to be a very heterogeneous granitoid body, further dating of the major phases of this Pluton should only be attempted after their intrusive and genetic relationships are better understood.

The Port Mouton Pluton is of interest because of its well exposed contacts with the Meguma, its extreme heterogeneity, its beryllium potential and as a key to the further understanding of Devonian granites in Nova Scotia.

LITHOLOGIES

The predominant granitic type in this Pluton is a medium grained biotite-muscovite mesocratic granitoid (Unit 4, described below). It rarely contains large feldspar phenocrysts that are characteristic of larger plutons or batholiths. In places, the Unit develops distinct banding. Along the eastern exposure of the Pluton, this banding appears to be relatively consistent (217/74°NE). The bands consist of a mesocratic

phase (Unit 4) with a more leucocratic phase (Unit 3) depleted in biotite.

The mappable units in the western block of the Pluton (west of South West Port Mouton) are listed in chronological order from oldest to youngest.

UNIT 1

Coarse grained biotite mafic granitoid (tonalite) coarse grained biotite-muscovite mesocratic granitoid (granodiorite). Both of these granitoids appear to be the oldest granitoid phases in the Pluton. Nowhere do these two units appear in the same outcrop. It is probable that the tonalite grades into the granodiorite. The tonalite and the granodiorite are equigranular, locally subporphyritic, granitoids. Both the tonalite and the granodiorite contain inclusions of the country rock (pelite and psammite). The degree of assimilation and size of these xenoliths varies as a function of distance, within the country rock, from the contact.

UNIT 2

Coarse grained biotite-muscovite leucocratic granitoid. This unit is equigranular and quartz rich. In places it contains biotite schlieren (up to 3% of the outcrop). This is a relatively minor unit. It is clearly younger than Unit 1, as it contains inclusions of this unit. Its relationship to Units 3 to 8 is unclear to date.

UNIT 3

Medium grained biotite-muscovite leucocratic granitoid to medium grained leucocratic granitoid with minor pegmatite. This unit is equigranular. It contains quartz, plagioclase and potassic feldspar with traces of biotite and muscovite.

UNIT 4

Medium grained biotite-muscovite mesocratic monzogranite to a medium grained biotite mafic granitoid. It is unclear where the transition between these subunits occurs. Both the monzogranite and the granitoid are equigranular. The contacts between Units 3 and 4 are indistinct in places. Well-developed interbanding between Units 3 and 4 occurs in the eastern sector of the Pluton. Inclusions of Unit 3 occasionally occur in the medium grained biotite-muscovite mesocratic units. The inclusions are subrounded and can have indistinct wispy contacts with the surrounding granitoid. The size of the inclusions ranges from approximately 5 cm to 30 cm. Occasionally the contacts between Unit 3 and Unit 4 are highly convoluted. Unit 4 contains inclusions of Unit 1, Unit 2 and the country rock. This unit becomes fine grained at some of the country rock contacts, making the distinction between it and Unit 1 and Unit 6 difficult in places.

UNIT 5

Fine grained biotite (minor muscovite) mafic granitoid breccia. This unit contains up to 50%

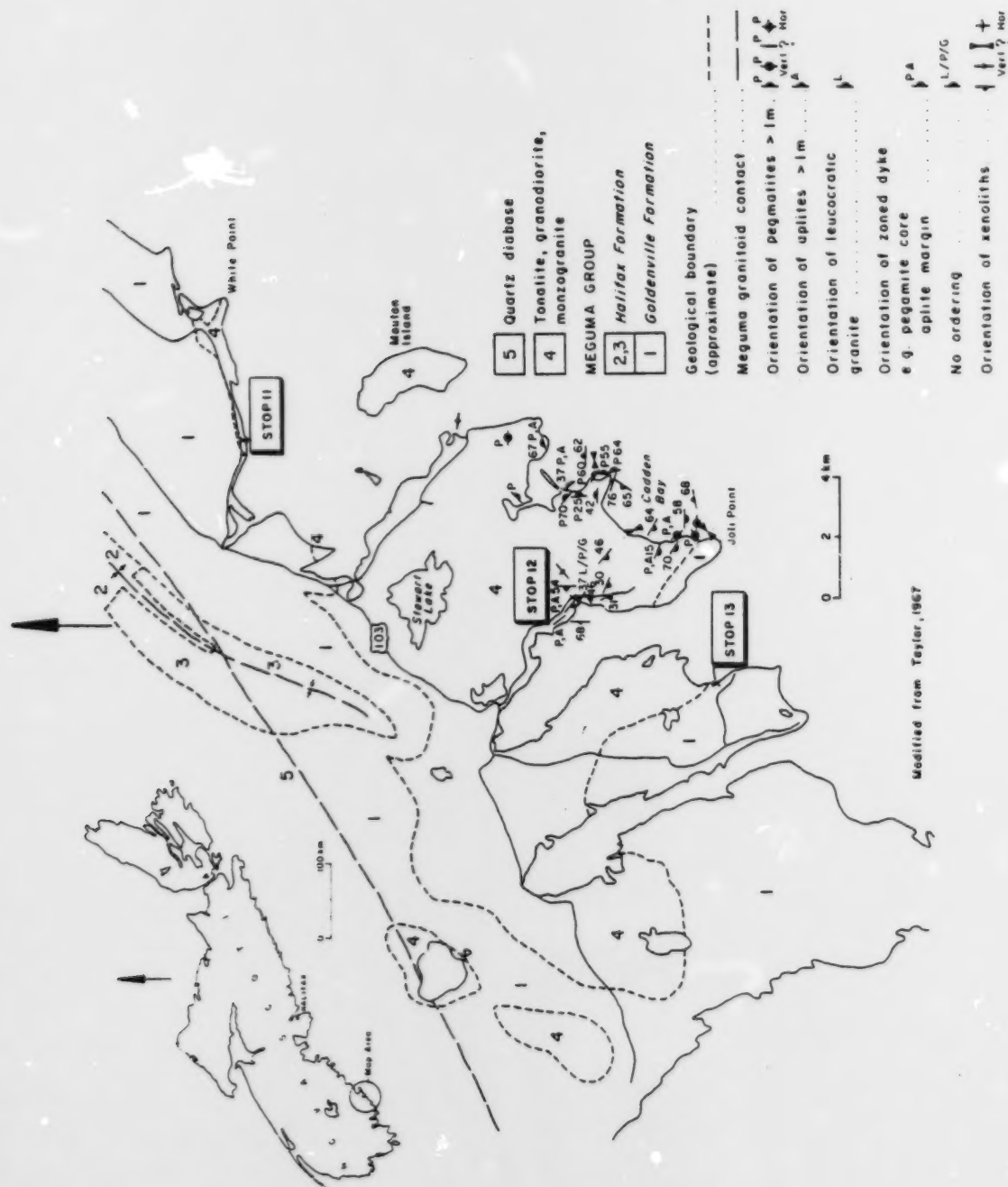


Figure 6-1. Map showing the geology and the field trip stops in the area of the Port Mouton Pluton.

angular to subrounded 'clasts' of quartz lenses, country rock, Unit 1 (tonalite) and Unit 2 (medium grained leucocratic granitoid). Clasts range in size from 5 cm to approximately 50 cm.

UNIT 6

Fine grained biotite (minor muscovite) mesocratic to mafic granitoid. This unit appears as dyke-like bodies. It is equigranular to locally subporphyritic. The type of inclusions (and percentage of inclusions) vary. Rounded biotite-rich 'xenoliths', medium grained leucocratic granitoid and rounded psammites are typical inclusions found in this unit. Inclusions of country rock (psammite) can reach up to 70 cm in diameter. The percentage of inclusions in this unit vary from zero to approximately 25%.

UNIT 7

Fine- to medium-grained biotite-muscovite leucocratic granitoid to a fine- to medium-grained leucocratic granitoid. This unit is equigranular and contains quartz, plagioclase, potassic feldspar and occasional garnet (when associated with pegmatites). In places, this unit contains inclusions of coarse grained biotite mafic tonalite (Unit 1) and country rock fragments. Some fine- to medium-grained biotite-muscovite leucocratic granitoid dykes are distinctly banded. This unit is often associated with zoned pegmatite dykes.

UNIT 8

Fine grained garnet-aplite-quartz-muscovite-feldspar pegmatite. The pegmatites vary dramatically in grain size throughout the Pluton. Single quartz crystals of up to one metre in length have been found. Traces of garnet and beryl are found occasionally in pegmatites associated with the granitoid/country rock contacts. Pegmatites often are unzoned and have no associated aplites.

The eastern section of the Port Mouton Pluton (along the coastline from New Building Cove to Port Mouton Head) is more complex than the western section. In a single 10 m² outcrop, up to six granitoid phases can be mapped. The complexity of this section of the Pluton is characterized by a shoreline outcrop located north of Davenport Point. The units listed from oldest to youngest are as follows:

- (a) coarse grained biotite mesocratic tonalite (Unit 1);
- (b) fine grained leucocratic granitoid (Unit 3?);
- (c) medium grained biotite-muscovite mesocratic granitoid (Unit 4);
- (d) fine grained biotite-muscovite mesocratic granitoid, with 20% rounded 5-20 cm psammite inclusions (Unit 6?);
- (e) fine grained biotite-muscovite mesocratic granitoid with no inclusions (clearly younger than (d) (Unit 6?);
- (f) aplite/pegmatite (Unit 8).

A new unit that is distinct to this section of the eastern coastline is a medium- to

coarse-grained biotite-muscovite mesocratic equigranular granitoid. This unit contains 30-100 cm rounded inclusions of fine grained biotite-muscovite mesocratic granitoids (Unit 6). This new unit, although similar in appearance to Unit 4 (of the western sector of the Pluton) is not the same as Unit 4. This unit contains inclusions of Unit 6 and is therefore younger than Unit 4.

CONTACT RELATIONSHIPS WITH THE MEGUMA

The contacts of the Pluton with the Goldenville Formation appear to be sharp in most places. Exceptions occur near Hunts Point and White Point and the southern contact along Cadden Bay. Migmatites are exposed at White Point. At Hunts Point thin dykes of medium grained biotite-muscovite mesocratic granitoid (Unit 4) invade the Goldenville Formation approximately parallel to bedding (039/71°NW). The dykes vary in width from five cm up to one metre wide and, in places, the Goldenville bands have become pygmatically folded. Aplite and pegmatite crosscut these granites. In the vicinity of Cadden Bay (along the shoreline) the contact between the Goldenville Formation and the intrusion is well exposed. Over a distance of approximately 2.5 km (north-south) there is a progressive change from thin granitoid dykes (or sills) invading the Goldenville along bedding planes (at Port Joli) to flow banded granitoids with occasional one to two metre long xenoliths of Goldenville at Harbour Rocks. At Port Joli Head, psammite (which comprises 90% of the outcrop) is invaded by a fine grained biotite-muscovite mafic granitoid (Unit 4 or a fine grained equivalent of Unit 1) along its bedding planes. To the north at an outcrop near Harbour Rocks, a coarse grained biotite-muscovite granodiorite (Unit 1) contains 5-10% ribbon shaped Goldenville inclusions.

The predominant trend of flow banding (defined by biotite schlieren and Goldenville inclusions) is relatively consistent along this 2.5 km stretch of shoreline (approximately 130/SW dip). Local perturbations occur at all angles to this regional trend. Occasional medium grained biotite-muscovite granitoid dykes (Unit 4) intrude the coarse grained biotite-muscovite granodiorite (Unit 1).

Extensive pegmatite dykes occur at most of the Meguma/granite contacts. The pegmatites are both zoned and unzoned. Contacts of the zoned pegmatites with the main medium grained biotite-muscovite granitoid (Unit 4) are often obscure. The pegmatites intrude both the Port Mouton Pluton and extend into the Goldenville Formation. The westernmost contact of the Pluton (along Highway 103) consists of a fine- to medium-grained muscovite-rich leucocratic cataclastic granitoid (Unit 3 or Unit 7). A shear zone (044-055/70°NE) appears to cross the Pluton in this region.

ECONOMIC POTENTIAL

The Port Mouton Pluton has been assessed for its beryllium, tin, tungsten and uranium potential. Beryllium exploration by the Nova Scotia Department of Mines in 1959 and by

McIntyre-Porcupine Mines Limited, indicate that the beryl associated with pegmatites is located near the contacts with the Meguma. Beryllium showings have been located at Sandy Cove, along Cadden Bay, Forbes Cove, New Building Cove and in the southwestern section of Port Mouton Island. Hexagonal shaped beryl (ranging from two to three mm to ten cm in length) are visible in these locations.

Uranium explorations appear to have been unsuccessful (Can-Lake Explorations Limited (1978)). Esso Minerals Canada (1981) completed geochemical sampling and reconnaissance geology of the Port Mouton areas in an attempt to locate tin and tungsten deposits. Tin anomalies in the Granite Brook and Wagner Brook areas were claimed to be promising.

ITINERARY

Travel from Liverpool southwest along Highway 103 to Exit 20 (Summerville Centre, Hunts Point, White Point). Turn left and drive 2.1 km/1.3 mi to a widened spot on the right shoulder of the road just past a blue house on the right.

STOP 11: CONTACT ZONE OF THE PORT MOUTON GRANITOID INTRUSION

The exposures extend along the shore from the parking spot to the blue house. They constitute part of an extensive contact zone between the Port Mouton granitoid Pluton and the Meguma country rock, now at a lower amphibolite facies grade to regional metamorphism. Some of the medium grained biotite-muscovite mesocratic granitoid (Unit 4) is clearly intrusive into the country rock along the bedding planes and shows a preference for the pelitic over the psammitic horizons. The pelites develop very biotite-rich margins in reacting with the magma and these types of biotite-rich inclusions can be seen in many phases in the interior of the Pluton. Other phases crosscutting the Meguma at high angles include both aplite and pegmatite.

Some of the psammitic horizons have clearly become ductile and show evidence of partial melting in the formation of pygmatically folded granitic bands which are probably partial melts of the Goldenville Form. These 'sweats', as well as the bulk psammites themselves, their calc-silicate pods, and the pelites all represent possible contaminants of the Port Mouton Pluton and contributors to geochemical variation in it.

Return to Highway 103 and turn left towards Shelburne. Travel along this highway for a distance of 10.6 km/6.6 mi and turn left on Port Joli Harbour road. Travel down the Port Joli road for a distance of 4.5 km/2.8 mi and stop at a right bend in the road where a sloping outcrop face can be seen on the left side of the road.

STOP 12: TYPICAL PORT MOUTON GRANITOID OUTCROP

The principal host rock here seems to be the coarse grained biotite mafic tonalite (Unit 1) which has inclusions of the Meguma country rock, as well as some questionable inclusions of a leucocratic granitoid phase. At the northeast

corner of the outcrop is a porphyritic phase (Unit 1?) with an uncertain relationship to the tonalite. They are both cut by a strongly banded aplite (Unit 7) containing inclusions of the tonalite host and the Meguma. Narrow pegmatites are also associated with this aplite.

This outcrop is typical of the complexity of phase relations in the Port Mouton Pluton where anywhere from three to six granitoid types can be observed in a single outcrop. One of the major problems is to recognize all of the lithologies in their various textural types and to correlate from one outcrop to another.

Return to Highway 103. Turn left and drive 1.6 km/1.0 mi to the Port LeHebert road. Turn left and travel down this road 4.5 km/2.8 mi, bearing to the left at all junctions until a small road on the left is reached. (If you come to the VanLoon farm property, you have travelled about 100 m too far). The outcrop is a ten minute walk down the small road to the shore.

STOP 13: BERYL-BEARING PEGMATITE

To date, the only economic interest in the Port Mouton Pluton has centered on the common occurrence of beryl-bearing pegmatites, most of which occur within the Pluton itself. Unfortunately, none of these is easily accessible. Thus we have chosen to look at this one which cuts only the Meguma, at least at this locality. The pegmatite itself is very wide and contains some garnet, numerous quartz-rich zones and a very few beryl crystals associated with the quartz. Fragments of the Meguma (psammitic and biotite schist) are included in this pegmatite complex. The xenoliths range in size from one to nine metres in length.

Please do not sample the beryl crystals in order that others may observe them in future.

CHAPTER 7

GRANITOID ROCKS OF SHELBURNE COUNTY AND
EASTERN YARMOUTH COUNTY, NOVA SCOTIA

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Rogers, H. D., Granitoid rocks of Shelburne County and eastern Yarmouth County, Nova Scotia; in Guide to Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

122 Granites & Mineral Deposits SW N.S.

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INTRODUCTION

Granitoid rocks of the southwestern corner of Nova Scotia (Fig. 7-1) have often been grouped with those of the South Mountain Batholith. Field and laboratory work by Taylor (1967, de Albuquerque (1977, 1979) and Rogers (in preparation) have made it apparent that major differences exist between these two groups of rocks. The objective of this part of the guidebook is to present some characteristics of these plutons and to point out the many ways in which they differ from the South Mountain Batholith.

GEOLOGICAL SETTING

The plutons, mostly of peraluminous composition, occur individually within amphibolite facies metasedimentary rocks of the Cambro-Ordovician Meguma Group which was folded and metamorphosed during the Acadian Orogeny. The contact of the Barrington - Passage Pluton is complex and irregular owing to migmatization of the country rock over a width of up to several kilometres. The contact of the Shelburne Pluton is commonly marked by large amounts of pegmatite. The other plutons appear to have sharp contacts. The parallel alignment of a tectonic foliation in the granitoid rocks and the country rock suggest that all plutons, except the Seal Island Pluton, were intruded prior to the cessation of Acadian Orogeny folding. Preliminary whole rock, Rb-Sr dating of the granite unit of the Shelburne Pluton (Rogers, in preparation) suggests a middle Ordovician time of emplacement.

Diabase (norite of Albuquerque, 1979) occurring in the Barrington Pluton tonalite at Murray Cove, and diorite occurring in the granite-granodiorite phase of the Shelburne Pluton at Birchtown, appear to be xenoliths of unknown origin.

PETROLOGY

The Shelburne Pluton (290 km²) is composed of granite, tonalite, granodiorite and aplite in decreasing order of abundance. The principal lithologies (Fig. 7-2a) do not occur as map units (Fig. 7-3). The tonalite granodiorite, granite and aplite are medium grained and equigranular except for a porphyritic granite unit (Fig. 7-3). Mineralogy of the granite and granodiorite is Qz + Ksp + Plag + Musc ± Garnet. The occurrence of garnet is restricted to aplite, its surrounding granite and porphyritic granite. Mineralogy of the tonalite is Plag + Qz + Biot ± Ksp.

The Bald Mountain Pluton (55 km²) is composed of monzogranite (Fig. 7-2b) which is medium grained and equigranular. Mineralogy of the Pluton is Qz + Ksp + Plag + Biot + Musc. Several small granite intrusions, compositionally and texturally similar to the above, outcrop in the western part of the area (Fig. 7-1).

The Barrington - Passage Pluton (475 km²) consists mostly of tonalite (Fig. 7-2c). Small amounts of quartz diorite and granodiorite occur along the margin. Texture is commonly medium- to slightly coarse-grained, seriate porphyritic but

medium grained porphyritic and equigranular textures also occur. Mineralogy of the Pluton is Plag + Biot + Qz ± Ksp ± Musc. It differs from the tonalite unit of the Shelburne Pluton mainly in being more biotite-rich and in containing a less calcic plagioclase.

The Lyons Bay Pluton (<5 km²) is unique in being the only pluton of the area to contain amphibole. Composition varies from quartz diorite to tonalite (Fig. 7-2d). Texture is medium grained, equigranular. Mineralogy of the Pluton is Plag + Qz + Biot + Hornblende.

The Seal Island Pluton (>100 km²) consists of monzogranite and syenogranite (Fig. 7-2e) which occurs as biotite granite and muscovite-biotite granite (Fig. 7-4). Texture is coarse grained and porphyritic, similar to rocks of the South Mountain Batholith. Mineralogy of the Pluton is Qz + Ksp + Plag + Biot ± Musc.

GEOCHEMISTRY

Major and trace element average composition and standard deviation for the plutons are presented in Table 7-1. Major element data are also presented as Harker variation diagrams (Fig. 7-5). Data for selected trace elements are plotted in variation diagrams (Fig. 7-6).

With the exception of the Seal Island Pluton, major and trace element data indicate a lack of well-defined differentiation trends either within or between the Plutons. Major element trends do not conform to the average calc-alkaline trend (Gribble, 1969) and compositional fields differ from those of the South Mountain Batholith (McKenzie and Clarke, 1975). Trace element abundances and trends differ from those of the South Mountain Batholith (Smith and Turek, 1976; Chatterjee and Muecke, 1982), even for the Seal Island Pluton.

The Barrington - Passage Pluton is unusual in possessing a wide range of SiO₂ contents (60.91 to 72.39%) and in having K₂O and F contents which result from a biotite content (x = 25%) that decreases with increasing SiO₂. Within the granites, the Bald Mountain Pluton is the most homogeneous and contains the highest concentration of many lithophile trace elements (i.e., Be, Li, Sn, U, Rb). Granodiorite of the Shelburne Pluton is geochemically (and visually) indistinguishable from the granite unit so they are combined in the data presentations.

PETROGENESIS

The occurrence of rather basic compositions in the Plutons (eg. tonalite and quartz diorite), recognition of the xenolithic nature (cognate?) of the diabase (Barrington) and diorite (Shelburne) bodies, and the lower oxygen isotope ratios in some of the Plutons (Longstaffe et al., 1980), suggest a component of mantle material ± lower crustal (granulite facies?) material in these Plutons. Plutonism might be explained by injection of basaltic magmas into the lower crust, followed by various degrees of fusion and hybridization of crustal materials to produce much of the observed variation in the intrusions.



Figure 7-1. General geology of the area.

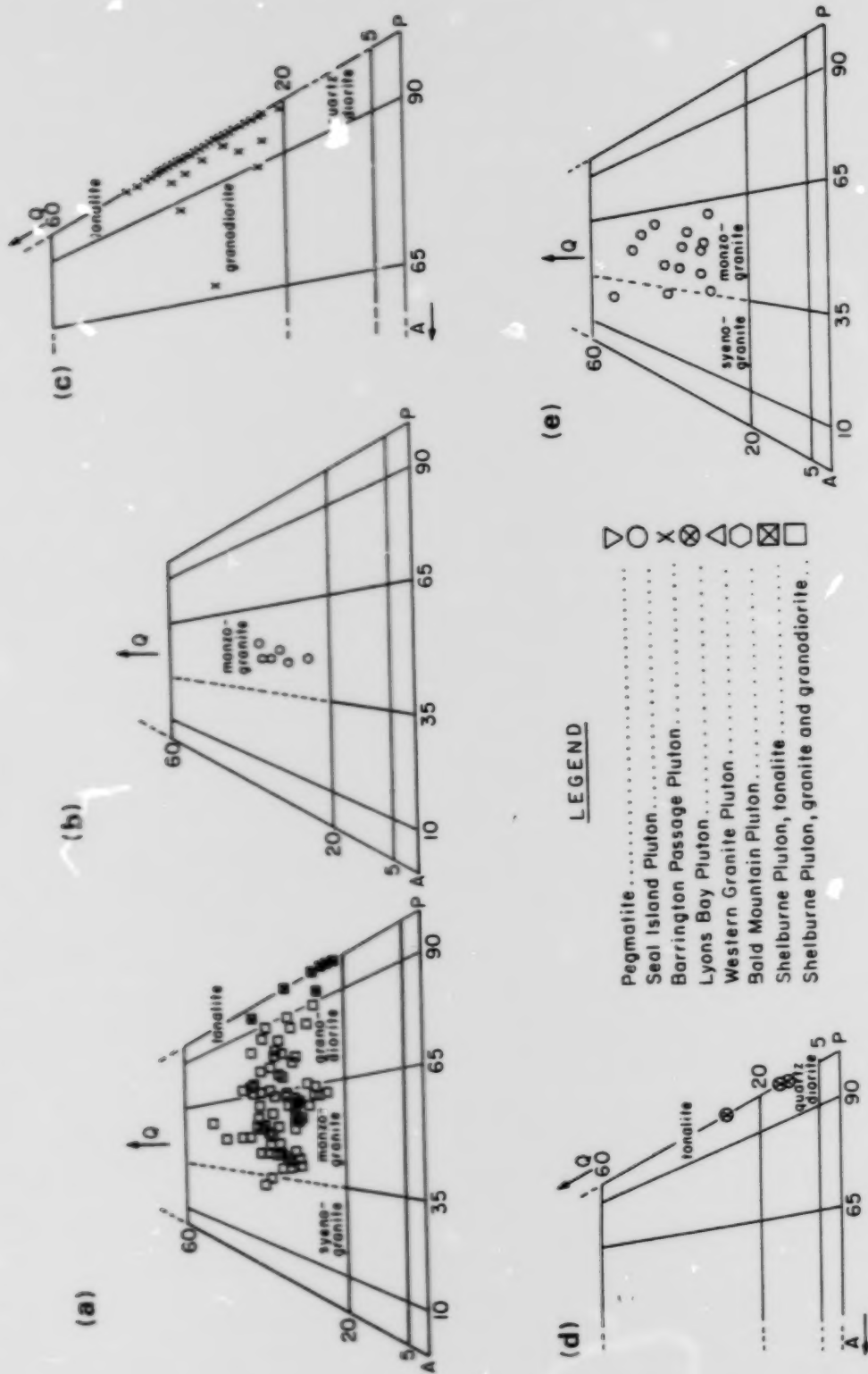


Figure 7-2. Modal composition of samples from the Plutons of the area (plotted after Streckeisen, 1976).

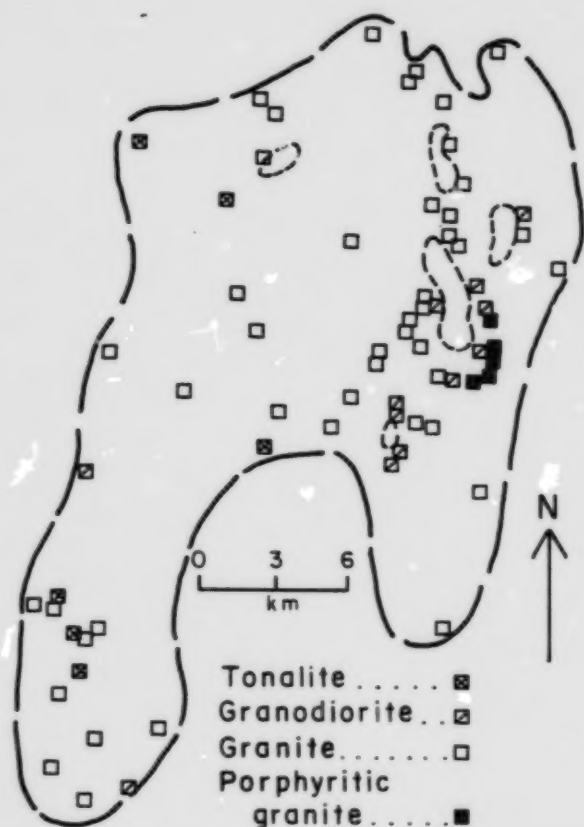


Figure 7-3. Distribution and composition of outcrops of the Shelburne Pluton.

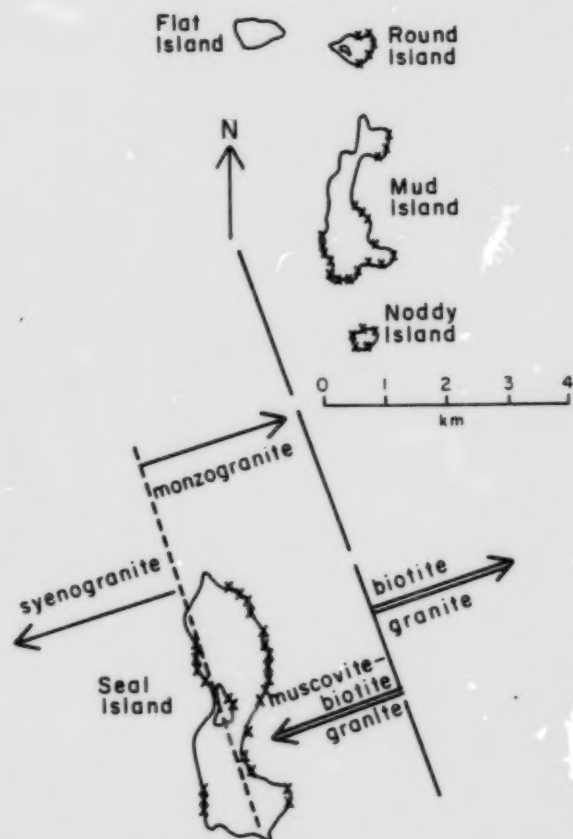


Figure 7-4. Distribution of rock types in the Seal Island Pluton.

Table 7-1. Major and trace element abundances and standard deviations of the Plutons of the area.

	Lyons Bay Pluton	Barrington Passage Pluton	Shelburne Pluton		Bald Mountain Pluton	Western Granite Plutons	Pegmatite	Seal Island Pluton
	(n=2)	(n=32)	granite (n=29)	tonalite (n=4)	(n=4)	(n=4)	(n=5)	(n=5)
SiO ₂	59.3 ±0.6	66.6 ±2.6	73.4 ±0.9	66.3 ±1.6	73.1 ±1.1	72.9 ±1.3	73.4 ±0.5	73.3 ±0.9
TiO ₂	1.0 ±0.0	0.7 ±0.2	0.2 ±0.1	0.5 ±0.1	0.2 ±0.0	0.3 ±0.1	0.0 ±0.0	0.3 ±0.1
Al ₂ O ₃	17.6 ±0.1	16.1 ±1.1	14.7 ±0.6	17.8 ±1.8	14.7 ±0.5	15.1 ±0.2	15.0 ±0.2	14.3 ±0.2
Fe ₂ O ₃ ^T	6.0 ±0.1	4.5 ±1.0	1.5 ±0.4	3.3 ±0.8	1.4 ±0.1	1.8 ±0.8	0.1 ±0.3	2.2 ±0.5
MnO	0.08±0.01	0.08±0.02	0.05±0.05	0.06±0.01	0.03±0.01	0.04±0.01	0.02±0.0	0.04±0.01
MgO	3.6 ±0.1	1.9 ±0.5	0.4 ±0.2	1.3 ±0.2	0.3 ±0.2	0.5 ±0.3	0.1 ±0.0	0.4 ±0.1
CaO	5.6 ±0.9	3.1 ±0.6	1.1 ±0.6	3.8 ±0.9	0.7 ±0.0	0.7 ±0.4	0.5 ±0.3	0.8 ±0.4
Na ₂ O	3.6 ±0.0	3.7 ±0.5	3.8 ±0.6	4.6 ±0.1	3.6 ±0.3	3.5 ±0.3	3.8 ±0.9	3.4 ±0.2
K ₂ O	2.1 ±0.5	2.3 ±0.4	3.8 ±1.0	1.6 ±0.3	4.5 ±0.0	4.4 ±0.4	5.7 ±1.9	4.4 ±0.2
P ₂ O ₅	0.4 ±0.0	0.2 ±0.1	0.2 ±0.1	0.1 ±0.0	0.3 ±0.0	0.3 ±0.1	0.2 ±0.1	0.2 ±0.1
A/(C+N+K)	0.92±0.04	1.13±0.07	1.19±0.09	1.08±0.01	1.23±0.02	1.28±0.07	1.13±0.05	1.22±0.08
Ba	362±54	513±266	531±165	389±100	459±115	601±183	96±150	592±95
Be		2.6±0.8	4.0±1.9	2.0±1.0	5.9±1.3	3.8±0.7	----	3.9±0.5
Cr		67±24	45±26	54±18	41±3	29±3	----	36±4
F		488±165	268±100	375±67	378±41	360±147	317±239	560±231
Ga		20±2	17±2	20±2	18±1	19±2	----	23±3
Li		58±3	85±48	71±22	15±15	61±44	14±7	63±24
Nb		11±4	8.8±1.9	6.0±0.8	9.3±1.5	11±1.4	----	19±1
Ni		14±6	5.2±5.2	5.8±3.8	2.3±1.2	5.0±1.6	1.6±1.1	7.8±3.1
Pb		16±5	23±7	11±2	26±4	26±3	4.7±1.2	19±3
Rb	73±25	87±25	142±51	55±5	195±19	188±77	238±121	166±22
Sn		2.9±1.2	4.6±1.4	2.1±0.7	5.8±2.0	4.6±0.8	2.2±2.2	4.5±1
Sr	373±11	324±127	146±109	414±246	62±5	97±34	44±56	125±53
Th		7.5±5.3	11±7	1.8±1.0	8.3±4.0	14±11	1.6±1.9	12±3
U		1.8±0.7	5.2±1.9	1.9±0.8	5.4±1.1	4.1±0.6	4.1±2.1	5.0±2.3
V		75±34	23±35	58±16	5.0±3.0	13±9	----	14±10
Y		18±8	16±4	10±2	12±3	17±6	----	21±5
Zn		65±14	49±18	54±3	44±5	55±5	15±15	51±10
Zr	132±30	173±51	107±55	136±18	70±7	110±44	12±13	160±33

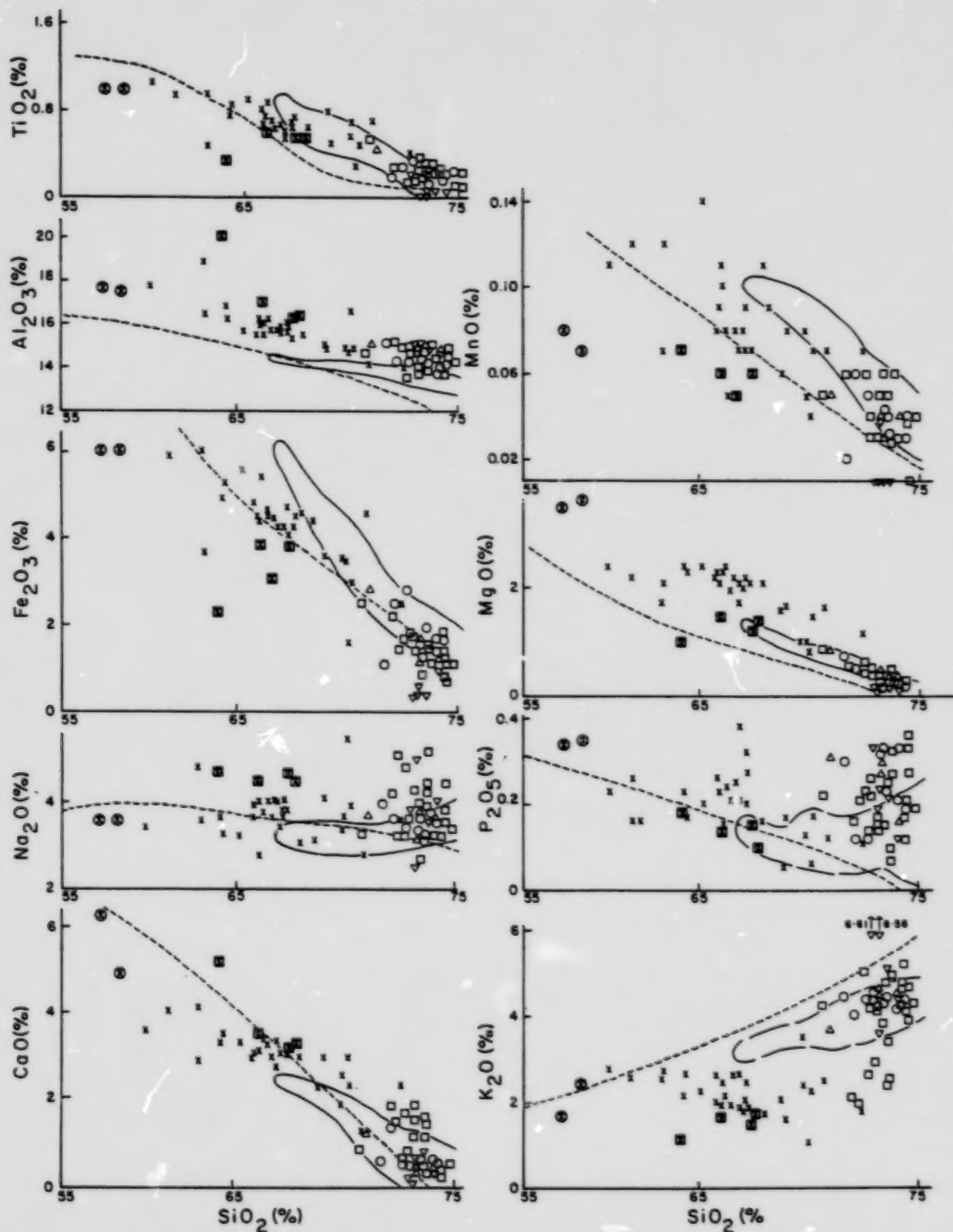


Figure 7-5. Harker variation diagrams for major element oxides. Dashed line is average calc-alkaline trend (Gribble, 1969). Solid line outlines field of South Mountain Batholith (McKenzie and Clarke, 1975). Legend same as in Figure 7-2.

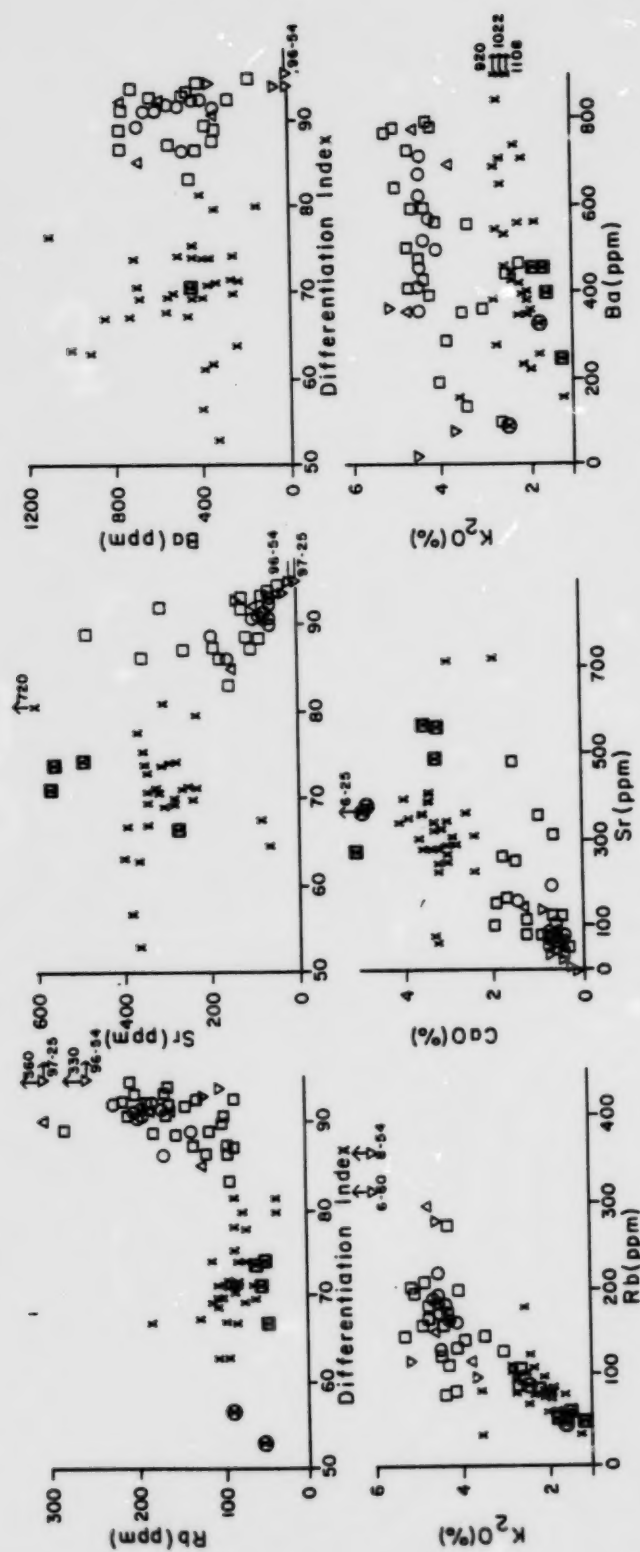


Figure 7-6. Abundance and distribution of selected trace elements.

The compositional and age differences set these Plutons apart from the South Mountain Batholith. Only the Seal Island Pluton appears to have formed at a later date during the magmatic event that formed the SM8.

ECONOMIC GEOLOGY

Mineral deposits and showings are absent from the area (Fig. 7-1). Present lithogeochemical data (Table 7-1) offer little encouragement that economic concentrations of these elements are present. The Seal Island Pluton is an exception in being a differentiated Pluton showing signs of hydrothermal activity.

ITINERARY

STOP 14: SHELburne PLUTON OUTCROP EXPOSED IN ROADCUT ON ROUTE 103, 5.5 KM WEST OF EXIT 26 TO THE TOWN OF SHELburne.

Buff to light grey, medium grained, equigranular muscovite-biotite granite containing rare, medium grained, garnetiferous aplite lenses and crosscut by unzoned pegmatite dykes comprise the outcrop. A tectonic foliation, readily detectable as biotite grain alignment, pervades the outcrop. Xenoliths of schist and quartzite parallel the foliation. Friable shear zones, having an average northeast trend, occur throughout the outcrop. A rare flow foliation is locally developed at the southwest corner of the outcrop. North-dipping slickensided surfaces occur at the southeast corner of the outcrop.

STOP 15: BARRINGTON PASSAGE PLUTON EXPOSED IN ROADCUT ON ROUTE 103, 2.6 KM WEST OF THE START OF ROUTE 103 AT OAK PARK

Blue-grey, slightly coarse grained, subophyritic biotite tonalite containing abundant biotite schlieren, crosscut by pegmatitic dykes, make up this outcrop. A tectonic foliation and friable shear zones, similar to those at Stop 14, are present. The pegmatite dykes differ from those of the Shelburne Pluton in being relatively narrow and a few are zoned. Pegmatitic dykes are commonly far fewer in outcrops of this Pluton.

CHAPTER 8

MAGMATIC FRACTIONATION AND METASOMATIC ALTERATION
OF THE DAVIS LAKE-EAST KEMPTVILLE COMPLEX
AND ITS RELATIONSHIP TO THE REST OF THE SOUTH MOUNTAIN BATHOLITH

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Richardson, J., Magmatic fractionation and metasomatic alteration of the Davis Lake East Kemptville Complex and its relationship to the rest of the South Mountain Batholith; in Guide to the Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

INTRODUCTION

The East Kemptville greisen-hosted tin mine, (58 million tonnes of 0.165% Sn) is located 300 km southwest of Halifax at 44°16'N and 65°41'W in Yarmouth County, Nova Scotia, Canada (Fig. 8-1, Index Map). It occurs in highly differentiated and altered rocks of the Davis Lake monzogranite, immediately below an inflection in its contact with metawacke of the flyschoid Meguma Group. In 1978, Shell Canada Resources Limited discovered mineralized boulders in the glacial till near Kemptville. This led to the delineation of an extremely stanniferous body that had been previously mapped at a reconnaissance scale as undifferentiated granitic rock. Shell outlined a mineralized zone 1500 m x 500 m x 100 m. Cassiterite is the main tin mineral present, with minor economic chalcopyrite and sphalerite, and economically insignificant amounts of wolframite. The property was purchased in 1982 by Rio Algom Limited.

Information for this study was obtained from 14 600 m of diamond-drill core, two trenches totalling 225 m in length, approximately 1.2 km of underground development and the five known outcrops primarily during nine months of field work in 1979-1980. In 1981, ten days were spent mapping underground in the decline. Shorter visits have been made annually since 1981.

This paper briefly summarizes the geological relationships of the primary igneous phases, albitic alteration zones, the greisen-hosted mineralization and later veins types and goes on to discuss the geochemistry of the Davis Lake Complex with respect to the South Mountain Batholith (SMB). More detailed work is in progress to better define the conditions responsible for the generation of the host rocks and the various alteration phases.

GEOLOGICAL SETTING

REGIONAL SETTING

The term "Davis Lake Pluton" was first used by McAuslan et al. (1980) for the granitic body defined by field relationships, geologic mapping and diamond drilling to lie in the vicinity of Davis Lake. Petrographic and geochemical evidence suggests that the least altered rocks are monzogranites (Richardson, 1983, in press). The term "Davis Lake Complex" is used here to refer to all granitic rocks in the Davis Lake area. Although poorly exposed, intrusive contacts in boulders and airborne radiometrics (R. Slaney, personal communication) suggest that complex intrudes the extreme southwest end of the Devonian-Carboniferous SMB, a 300 km x 60 km composite, posttectonic, peraluminous intrusion of biotite granodiorite (McKenzie and Clarke, 1975; Clarke and Muecke, 1980) that also hosts other areas of Sn-U-W mineralization (O'Reilly et al. 1982; Charest et al., this volume; Logothetis, this volume). Chatterjee et al. (1983) used a multivariate approach to distinguish between tin and uranium specialized granites within the Batholith.

The East Kemptville deposit occurs below the northeast contact of the Davis Lake Complex with

the overlying Meguma Group. This metasedimentary package is divided into three lithologic units: the metawacke that overlies the greisen zone, succeeded by a transitional zone of interbedded metawacke and meta-argillite, and meta-argillite (Fig. 8-1). These units occur in a series of closely spaced, northeast-trending folds.

GRANITIC ROCKS OF THE DAVIS LAKE COMPLEX

Although there is poor outcrop control, extensive drilling throughout the Davis Lake Complex indicates that it is likely a single intrusion extensively modified by hydrothermal activity.

Briefly, biotite monzogranite is the least altered rock type in the Complex. Characteristically, it contains approximately 5% biotite often partially altered to chlorite or sericite. Large euhedral plagioclase, quartz and microcline perthite phenocrysts occur in a medium grained groundmass of quartz, orthoclase and perthite. This rock type gradationally alters to blue leucocratic monzogranite as the biotite is totally replaced by chlorite or white mica (or both) aggregates. Plagioclase becomes small anhedral masses. Microcline perthite phenocrysts become subhedral. Cores of perthitic microcline are sericitized. Igneous exsolution textures are modified to "ganglia and patch" type albite (Richardson, 1983a, in press) characteristic of moderate albite-white mica alteration. Rims of secondary albite on perthite phenocrysts are common. Locally under the contact (usually near the deposit), there is a gradational change to the strong albite-white mica alteration that corresponds to variably altered leucocratic monzogranite. Microcline phenocrysts occur as remnant anhedral masses replaced by secondary albite. White mica occurs marginal to most grains and as aggregates completely pseudomorphing groundmass minerals (1983; synthesized and illustrated in Richardson, in press).

Greisen alteration overprints both of these rock types. It is recognized by its grey or green color, associated sulphide minerals and lack of feldspar (Fig. 8-2c,d,e) Richardson (1983, in press) gives a more detailed petrologic description of these alteration phases, their areal distribution and geochemistry. The extent and intensity of alteration is probably controlled by several factors including proximity to the granite-sediment contact, the shape of this contact, and porosity and permeability of the surrounding rock types.

EAST KEMPTVILLE TIN DEPOSIT

The deposit lies under the flattening of the contact shown in Figures 8-3 and 8-4 (plan and cross-section). Figure 8-5 is a computer-generated simulation of the subsurface topography of the granite-metasedimentary contact (viewed from 315°) as defined by drillhole intersections and Shell's preliminary level plans. To the west, mineralization extends into the metasedimentary rocks, but is below economic grade. Approximately 100 metres below the surface, the abundance of greisen zones decreases, and a zone of altered blue leucocratic monzogranite remains

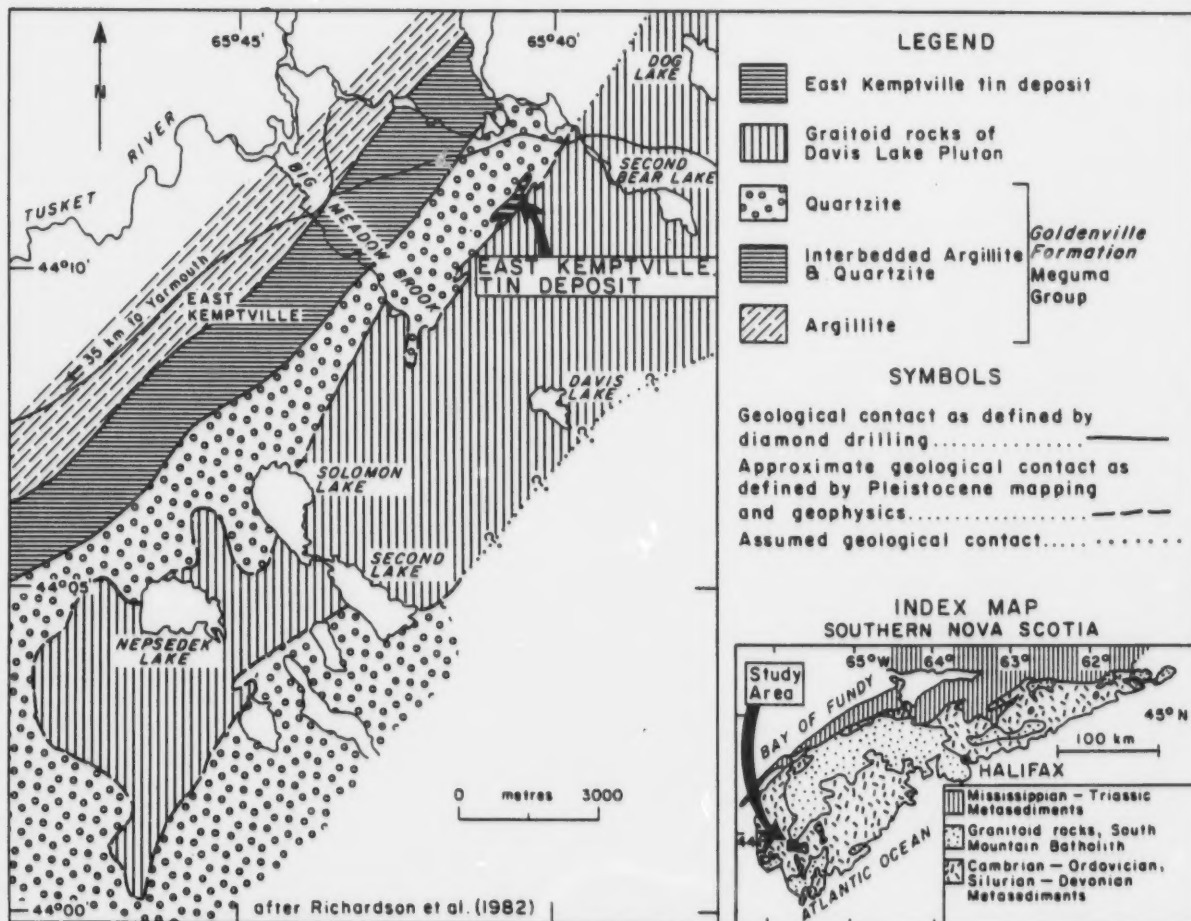


Figure 8-1. Geological map, East Kemptville greisen-hosted tin deposit, Davis Lake Complex, southwestern Nova Scotia.

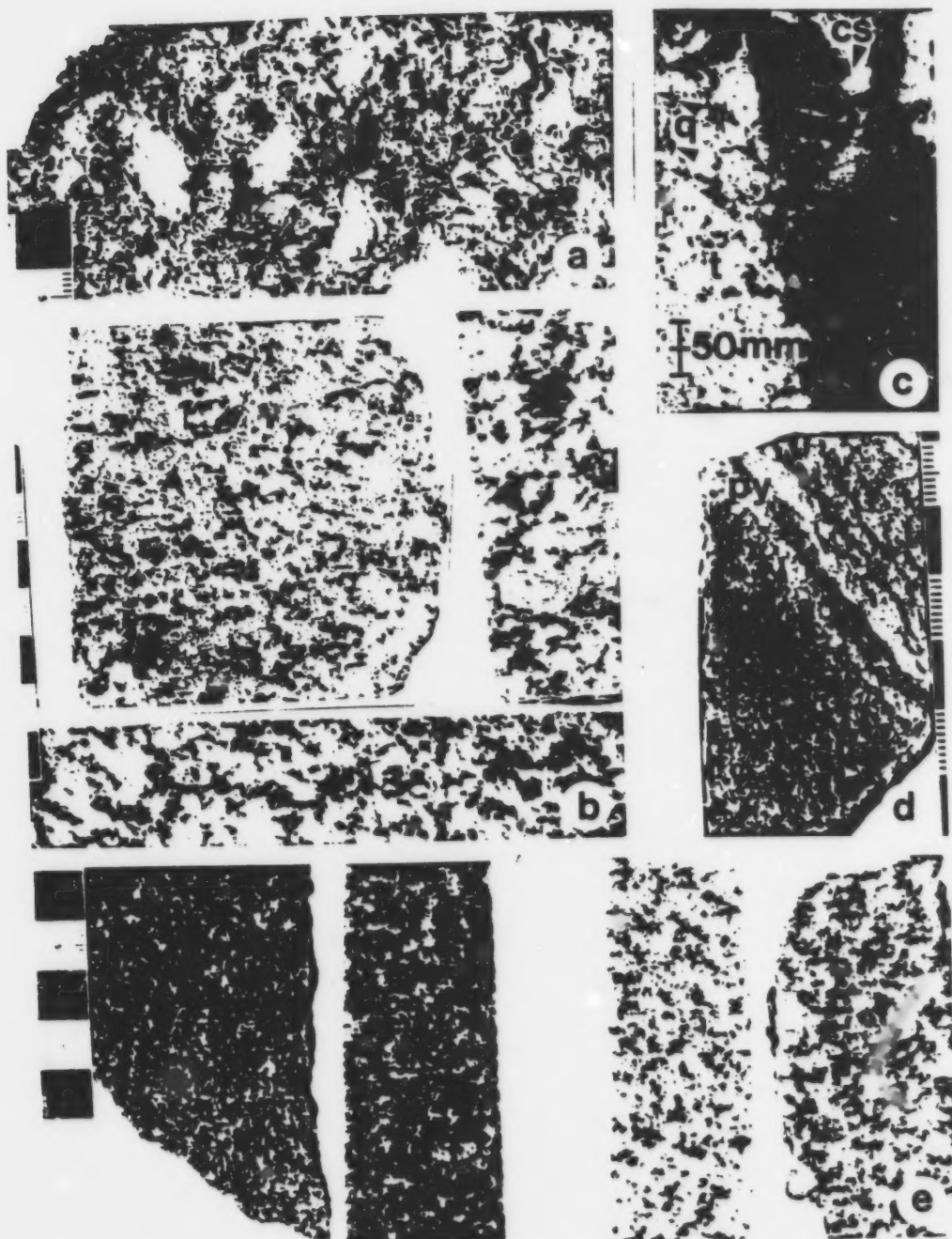


Figure 8-2. (a) Least altered biotite monzogranite, Davis Lake Complex. (b) Blue leucocratic monzogranite. Upper left sample stained for potassic feldspar. (c) Cassiterite-topaz core of greisen-bordered zone (cs-cassiterite, t-topaz, q-quartz) (d) Pyrite core flanked by topaz in center of greisen-bordered zone. (e) Dark and light massive greisen, East Kemptville tin deposit. Unless indicated, all scale bars are 1 cm. Core slabs (b,c,d,e) are approximately 2.5 cm wide.

[illegible]

Figure 6-4 Cross-section along L5800N, East Kemptville tin deposit.

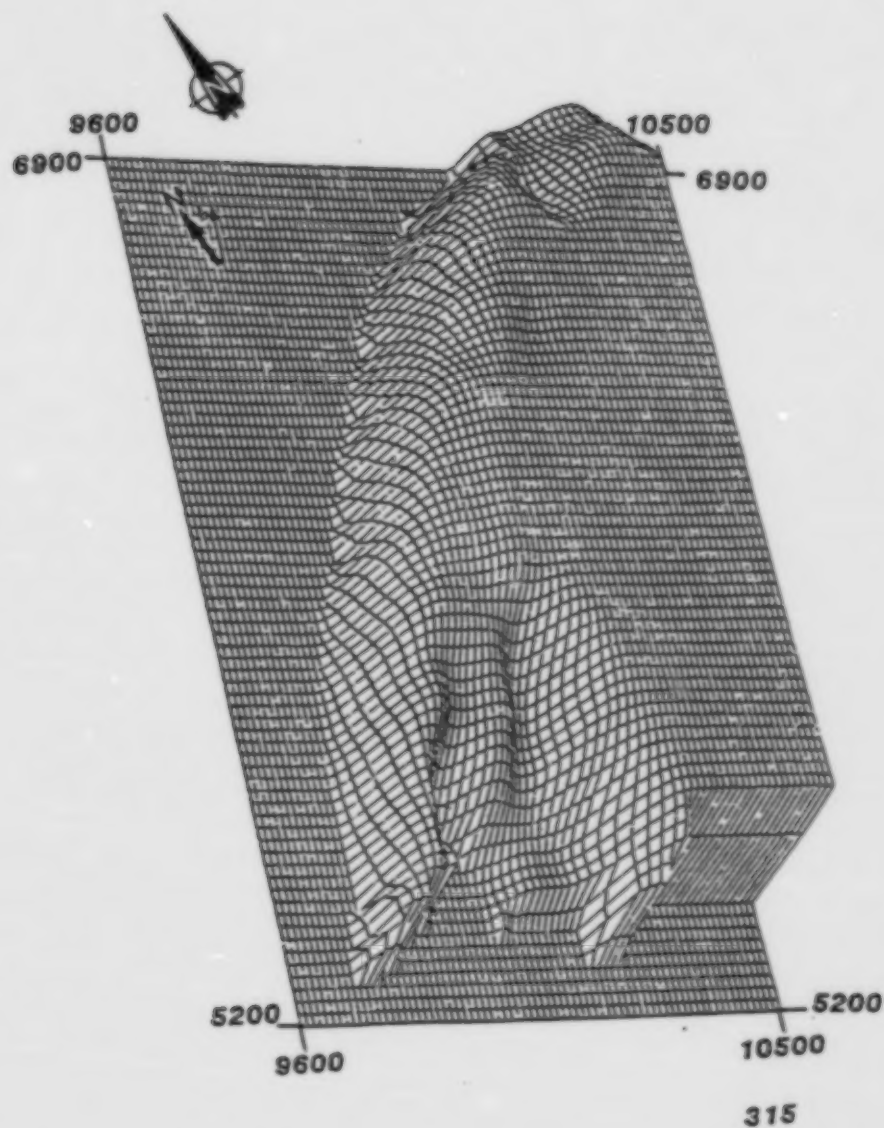


Figure 8-5. Computer-generated transect of surface of granite-metasedimentary contact, East Kemptville tin deposit. Viewed from 315°.

that contains less than 0.05% Sn. This cutoff grade can be traced to the surface and is used here as the eastern margin of the deposit. The northern and southern boundaries, also defined by a 0.05% Sn grade cut off, occur where the broad, shallow metasedimentary roof pendants coalesce with the main northwest dipping contact, as shown in Figures 8-3, 8-4 and 8-5. Metasedimentary rocks in the area of the contact are sheared, but the contact is sharp and well defined. Brecciation is limited. Granite dykes are restricted to a one to three metre zone near the contact. These observations suggest that, in the area of the deposit and along the northwest contact, the intrusion of the Davis Lake monzogranite, subsequent alteration and mineralization were relatively passive, unexplosive events.

Several thin (3-50 cm), fine grained, pre-greisen felsite dykes and pegmatite dykes and segregations crosscut the granitoid rocks within the deposit (Fig. 8-6). They cannot be correlated across drillholes or through the decline.

STYLE OF MINERALIZATION

The East Kemptville tin deposit consists of a stockwork of mineralized greisen-bordered microfractures and massive greisen zones located entirely within the Davis Lake monzogranite. The term "greisen" (German "greise" or "greisstein") has been used since the 18th century to describe a grey, stanniferous quartz-mica aggregate. Ferguson & Bateman (1919) define greisen as an alteration product of granite consisting essentially of quartz and mica with varying amounts of fluorine- and boron-bearing minerals. Such alteration can occur as envelopes locally bordering microfractures and quartz veins. However, if small networks of fractures occur, large masses of rock can be altered. Essential minerals are quartz, mica, (muscovite or sericite or zinnwaldite), fluorite, topaz and tourmaline. Accessory minerals include cassiterite, pyrite, arsenopyrite, wolframite and sphalerite. In the broader sense of this definition, much of the Davis Lake Complex is pervasively greisenized or partly greisenized. However, in this study, the term greisen is used in a more restricted sense to describe those granitic rocks in which more than 90% of the primary and secondary feldspar has been replaced by the essential or accessory minerals mentioned above. Visible topaz or fluorite is usually apparent. "Incomplete greisen" is used when 50-90% of the feldspars are replaced. This type of alteration occurs in symmetrical envelopes around veins and fractures (Fig. 8-7) or as massive zones. The following description of the East Kemptville mineralization has been extracted from Richardson (1982, 1983, in press) for this guidebook. The described features are illustrated best in Richardson (in press).

In the East Kemptville deposit, cassiterite occurs as 0.1-5 mm blebs and crystals in the cores of zoned greisen alteration envelopes and in massive greisen zones that are related to microfracture stockworks (Fig. 8-2C). Both are structurally controlled, forming a branching system (Fig. 8-7 and 8-8). The essential mineralogy is similar, consisting of bluish

quartz, topaz and minor silverish or greenish mica. Accessory minerals include cassiterite, pyrite, sphalerite, chalcopryrite, pyrrhotite, arsenopyrite, molybdenite, wolframite and minor fluorite. Microfractures and veins with greisen borders have preferred orientations that are parallel and perpendicular to the contact (Fig. 8-9 and 8-10) respectively, and dipping slightly to the west and subhorizontally respectively.

Greisen envelopes vary in total width from 0.2 to 50 cm. They enclose a central microfracture or vein of variable composition with diffuse margins up to 2.5 cm wide that is symmetrically flanked by zoned alteration envelopes averaging four centimetres on one side. This central vein may consist of bluish quartz-cassiterite, fine grained sphalerite-pyrite-cassiterite, arsenopyrite or wolframite (or both), topaz-cassiterite-chalcopryrite-sphalerite or crystals of cassiterite or pyrite (or both) along fractures (Figs. 8-2c,d, and 8-7). The mineralogy of these veins can change horizontally or vertically over very short distances, so crosscutting relationships do not indicate consistent relative chronologies. Greisen-bordered microfractures or veins crosscut the felsic dykes described earlier. Well-defined greisen-bordered zones are more than 20 metres in length and appear to terminate gradually. The total width of the vein and intensity of alteration also diminishes gradually.

Massive greisen occurs in altered regions commonly 20 m wide that show intense greisenization and substantial enrichment in cassiterite. It is subdivided by color and mineralogy. Dark greisen is enriched in topaz, cassiterite, chalcopryrite, sphalerite, pyrite, lesser arsenopyrite, molybdenite and wolframite. It is dark grey with a dull, waxy luster. Light greisen contains quartz, silverish and greenish micas, less topaz, less cassiterite, chalcopryrite, pyrite and rare arsenopyrite or wolframite. It is lighter grey and has a vitreous luster (Fig. 8-2e). Green greisen is found more rarely and consists of green mica, cassiterite, sphalerite and fluorite. Topaz-cassiterite-chalcopryrite-sphalerite veins and aligned blebs of cassiterite and topaz occur throughout these zones and are probably related to microfracturing. In addition, greisen-bordered microfractures may trend into or out of such highly altered zones.

PARAGENETIC SEQUENCE

The complex paragenetic succession presented in Table 8-1 was established by trench mapping, core logging and underground mapping, followed by petrography, X-ray powder diffraction and microprobe analyses. Although each major stage can be distinguished on the basis of mineralogy, associated alteration and crosscutting relationships, there is commonly slight overlap. The mineralogy of the later vein stages is described in Richardson et al., (1982), Mandarino et al., (1983), Richardson (1983) and is fully illustrated in Richardson (in press).

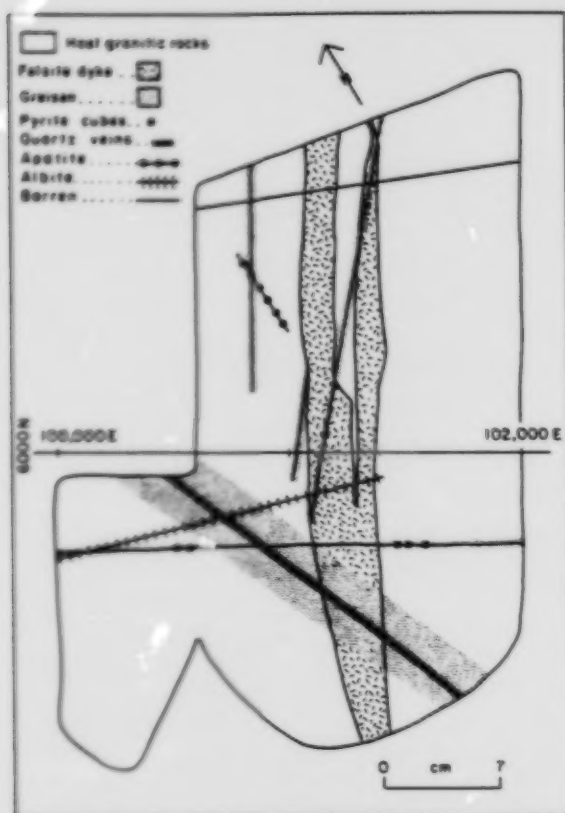


Figure 8-6. Map of small segment of Trench 6 on L6000N, East Kemptville. This segment illustrates the crosscutting relationships among the felsite dykes, veins or microfractures with greisen borders and several later vein groups.

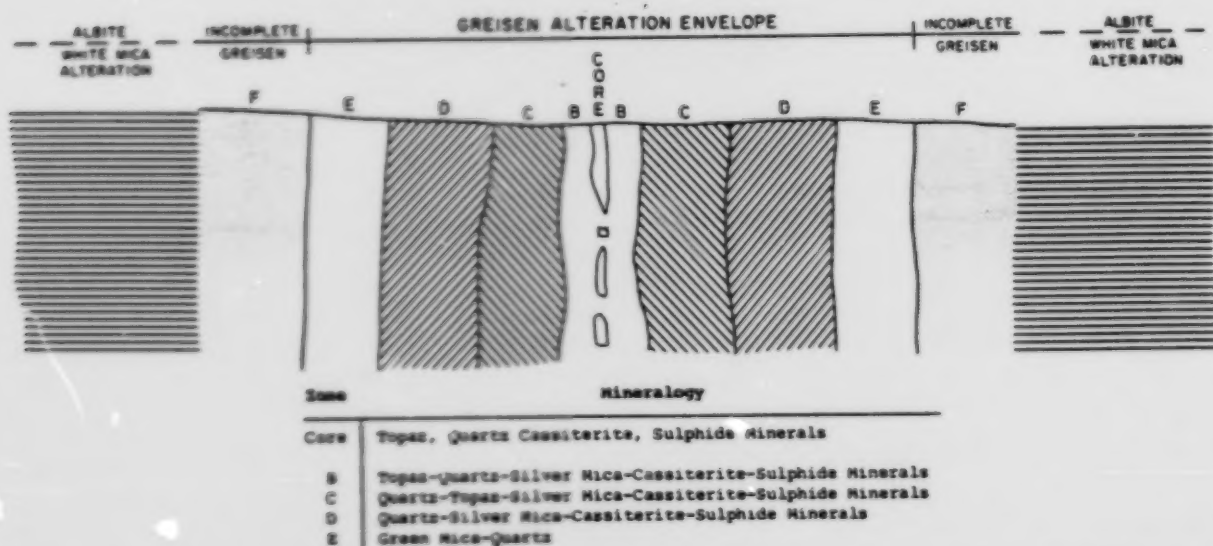


Figure 8-7. Schematic illustration of greisen alteration, East Kemptville. All zone boundaries are gradational.

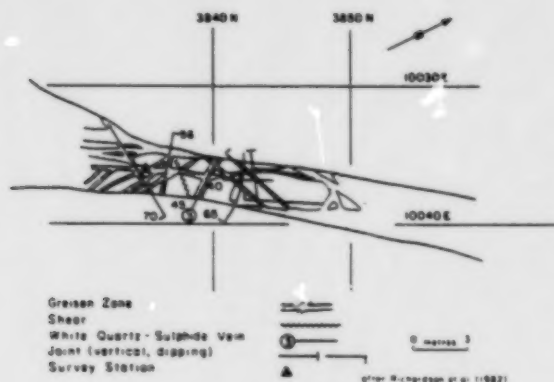


Figure 8-8. Segment of decline showing the oriented, branching network of greisen-bordered veins and microfractures.

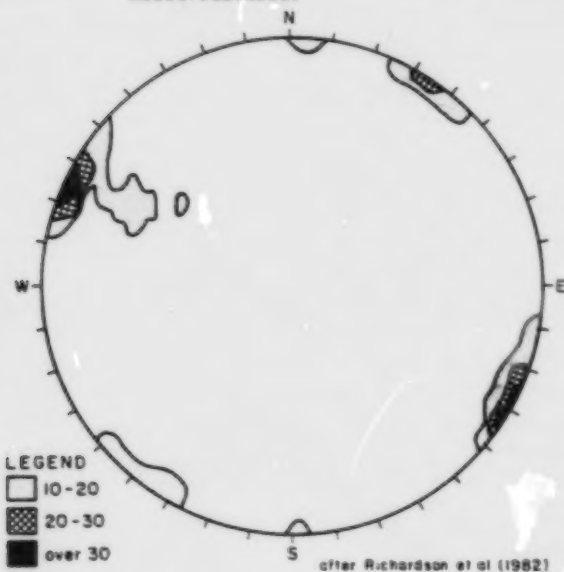


Figure 8-9. Rose net showing the strikes of 752 greisen alteration zones measured on surface and in the decline.

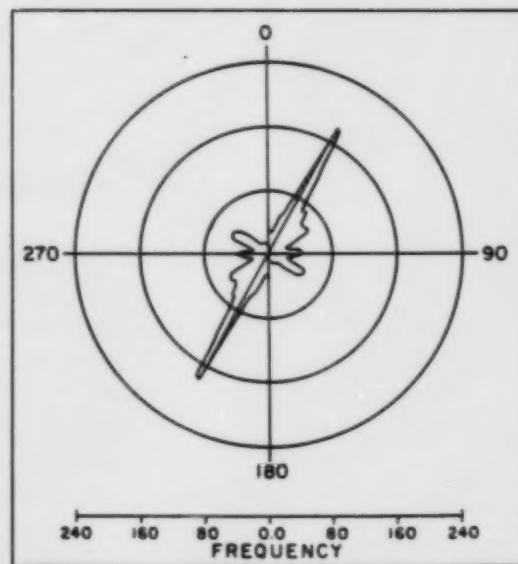


Figure 8-10. Contoured lower hemisphere projection of 256 poles to the plane of greisen-bordered veins measured on surface in the two trenches.

Table 8-1. Paragenetic sequence, East Kemptville tin deposit.

	Massive Greisen	Aspy WP	Greisen-Bordered Microfractures Ca-Py Topaz Sulphide	White Quartz +Sulphide+Phosphate Sulphide Sulphide Phos Phos	Mineral Veins	Quartz-Carbonate +Fluorite	Barren Quartz	Clay Fluorite	Crustal Crystals
Cassiterite									
Chalcopyrite									
Sphalerite									
Pyrite									
Pyrrhotite									
Arsenopyrite									
Wolframite									
Molybdenite									
Galena									
Topaz									
Fluorite									
Quartz									
Green Muscovite									
Silver Mica									
Stannite									
Fluorapatite									
Triplite									
Illite									
Siderite									
Dolomite									
Marcasite									
Dickite									
Stilbite									
Childrenite									
Vivanite									
Rhodochrosite									

after Richardson et al. (1982)

SAMPLING AND ANALYTICAL METHODS

Samples were collected from diamond-drill core as authorized by Shell Canada Resources Limited, from trenches within the deposit and the few available outcrops. Due to the medium to coarse grain size, large samples were taken to ensure that the subsequent analyses were representative. Outcrop samples weighed 1.5-2.0 kg. Core samples weighed slightly less, but represent a 2.5-3.0 m interval. The data presented here are the result of a pilot study done to determine the geochemical profiles of least altered, biotite monzogranite, the alteration assemblages and the greisen mineralization. Chatterjee and Strong (1984) and Boyle (GSC-in progress) provide a more comprehensive suite of analyses.

For this study, samples were reduced to 0.75 cm² using a jaw crusher. A 120 g split was ground to -200 mesh in a ceramic swing mill. Whole rock, major element analyses were done by X-ray fluorescence (XRF) at X-ray Assay Laboratories Limited (XRAL) Toronto, Ontario. FeO was determined at XRAL by wet chemistry. Fe₂O₃ was calculated. Nb was analysed at XRAL by XRF. F was determined by selective ion electrode at XRAL. Y and Zr were analysed by the author at University of Toronto by XRF. The analytical procedures outlined by Barnes and Gorton (1984) were followed for INAA of ¹³⁹La, and for rare earth element analysis using the Slowpoke II nuclear reactor at the University of Toronto. Since preliminary results for variably altered monzogranite and greisen were near the detection limit, the sample size was doubled (0.4 g) and counting time tripled (6-8 hrs.). For further information concerning standards, precision and accuracy, see Richardson (1983).

For the following discussion, geochemical data for the South Mountain Batholith were gathered from McKenzie and Clarke (1975), Charest (1976), Farley (1978), Allen and Clarke (1980), Clarke and Muecke (1980), Muecke and Clarke (1981), Clarke and Halliday (1980), and Logothetis (1984). To avoid repetition, analyses, descriptions or other information from this group will be referenced as Clarke and coworkers (1975-1985).

GEOCHEMISTRY AND NORMATIVE MINERALOGY

Although care was used in selecting the most unaltered samples for chemical analyses, few rocks in the Davis Lake region have completely escaped interaction with hydrothermal fluids. The freshest, least altered biotite monzogranite is peraluminous in the sense of Shand (1927), i.e. strongly corundum normative (2.01 av.) and contains normative anorthite (Table 8-2). It is more siliceous and contains less Al₂O₃ than the average granitic rock of LeMaitre (1976), the barren, precursor and specialized granites of Tschendorf (1977), and the granitic rocks of the South Mountain Batholith (Clarke and coworkers, 1975-1985). It contains approximately the same amount of TiO₂ and MgO as monzogranites of the SMB, and precursor and specialized granites. The average granite of LeMaitre (1976), barren granites and the granodiorite of the SMB are significantly

higher in these elements (1.31-0.63 wt%). In all the other rock types, CaO is two to five times higher than that of the Davis Lake biotite monzogranite. Average K₂O/Na₂O is similar. The P₂O₅ content of the Davis Lake biotite monzogranite is similar to the SMB, as both are two times greater than the average granites of LeMaitre (1976). Fluorine contents are similar to those from the New Ross Vaughan Complex (NRVC) (Table 8-3). Overall, the Davis Lake biotite monzogranite is very similar to the other monzogranites of the SMB as described by Clarke and coworkers (1975-1985), and within the range of Tschendorf's (1977) granites specialized for lithophile element mineralization.

Within the Davis Lake Complex, blue leucocratic and variably altered leucocratic monzogranites (Table 8-2) are slightly less siliceous than the biotite monzogranite. TiO₂ and MgO are lower and Al₂O₃, F and P₂O₅ marked? higher, Na₂O and CaO slightly higher in the altered rocks. K₂O is constant. The Fe₂O₃/FeO ratio is lower. Increasing normative apatite and corundum and decreasing quartz correspond to the intensity of albite-white mica alteration observed petrographically in the Davis Lake Complex. Normative anorthite is low or does not occur in altered rocks marginal or within the deposit. The major chemical components of greisen are SiO₂, Al₂O₃, Fe₂O₃, FeO, F, S and SnO₂ (Table 8-2). It appears to be less siliceous than the associated leucocratic monzogranites.

MAGMATIC FRACTIONATION VS. HYDROTHERMAL ALTERATION

Early descriptive geology and mapping of the SMB (Taylor, 1967, 1969, Smitheringale, 1973) is now complemented by the petrochemical studies of Chatterjee and Muecke (1982), Chatterjee et al., (1983), Clarke and coworkers (1975-1985), Longstaffe et al., (1980), most of which is reviewed in this volume. These studies, most particularly the major element chemical variations defined by McKenzie and Clarke (1975) and Rb/Sr isotope systematics discussed by Clarke and Halliday (1980), indicate that the monzogranites were derived from the biotite granodiorite by closed system fractional crystallization of biotite and plagioclase, i.e. the progressive differentiation of a peraluminous parent magma. However, this process alone cannot explain REE and other chemical differences between the monzogranites and leucomonzogranite (Muecke and Clarke, 1981). These differences and the widespread evidence of metasomatic activity (greisen, albitic, chloritic, fluoritic, hematitic and episyenitic alteration (Logothetis, 1984) and polymetallic lithophile and base metal mineralization throughout the Batholith strongly suggest that fluid-melt-rock interaction was also important during the late stage evolution of the Batholith (Clarke and coworkers, 1975-1985; Chatterjee et al., 1983; Logothetis, this volume).

The Davis Lake Complex was not sampled during much of this regional characterization. However, the extensive documentation of magmatic

Table 8-2. Whole rock, major element chemistry of Davis Lake Complex compared to that of average granitic rock (LeMaitre, 1976). Tischendorf's (1977) granitic classification and data for the South Mountain Batholith.

	Davis Lake Complex				LM	Tischendorf (1977)			SMB	
	BMG	BLM	VLM	GSN		B	P	S	GD	MG
SiO ₂	75.9	75.3	74.3	68.6	71.30	70.82	72.70	73.38	69.3	73.40
TiO ₂	0.20	0.00	0.06	0.03	0.31	0.34	0.24	0.16	0.63	0.24
Al ₂ O ₃	12.7	13.8	15.0	15.1	14.32	14.33	14.08	13.97	14.86	13.74
Fe ₂ O ₃	0.48	0.25	0.27	2.43	1.21	1.31	0.48	0.00	3.79	0.15
FeO	1.3	1.0	1.0	3.1	1.64	1.78	1.35	1.10	3.52	1.71
MnO	0.04	0.06	0.07	0.02	0.05	0.06	0.05	0.04	0.10	0.05
MgO	0.4	0.1	0.1	0.1	0.71	0.81	0.45	0.47	1.09	0.35
CaO	0.37	0.49	0.46	0.49	1.84	1.89	1.04	0.75	2.05	0.67
Na ₂ O	3.4	3.7	4.0	0.1	3.68	3.44	3.22	3.20	3.16	3.3
K ₂ O	4.25	4.23	3.90	0.28	4.07	4.34	4.84	4.69	4.00	5.0
P ₂ O ₅	0.10	0.27	0.72	0.33	0.05	*	*	*	0.16	0.12
n=	8	5	5	2	2460	2327	226	962	36	11

CIPW NORMATIVE MINERALOGY

qz	38.24	36.41	36.51	28.97	28.40	32.23	34.48	28.16	29.53
cor	1.81	2.80	5.21	0.64	0.54	1.68	2.30	1.93	1.15
or	25.21	25.25	23.10	24.25	25.88	29.06	28.12	23.03	29.78
ab	23.72	31.73	33.40	31.39	29.36	27.67	27.47	26.04	32.40
an	1.93	0.74	---	8.88	9.46	5.24	3.77	8.89	2.57
hy	2.59	1.88	2.04	3.39	3.79	2.29	2.37	5.06	3.61
mt	0.70	0.37	0.11	1.77	1.92	0.71	1.18	5.35	0.22
il	0.38	0.15	0.11	0.59	0.65	0.46	0.31	1.16	0.46
ap	0.24	0.64	1.70	0.12	*	*	*	0.37	0.28

Davis Lake Complex - all data from Richardson (1983a): BMG (biotite-monzogranite), BLM (blue leucocratic monzogranite), VLM (variably altered leucocratic monzogranite), GSN (greisen).

LM analyses of "average granite" (LeMaitre, 1976).

Tischendorf (1977) classification: B (barren granite), P (precursor granite), S (specialized granite).
* No data.

SMB (South Mountain batholith): GD (granodiorite) data from McKenzie and Clarke (1975), Smith (1974), Logothetis (1984), MG (monzogranite) data from McKenzie and Clarke (1975), Logothetis (1984).

Table 8-3. Abundances of fluorine and trace elements commonly used to indicate fractionation in the South Mountain Batholith and the Davis Lake Complex. All results in ppm unless indicated, code as listed in Table 8-2, including LMG (leucocratic monzogranite).

	South Mountain Batholith				Davis Lake Complex			
	GD	MG	LMG	GSN	BMG	BLM	VLM	GSN
n=	36	28	20		8	5	5	2
Ti	3780	1440	482	320	1125	472	360	210
Zr	172	74	50	48	112	38	21	20
Nb					20	30	40	20
Ta	1.94	2.86	6.83	6.34	0.9	2.2	3.1	2.9
Y					38	17	10	3
F	700	1500*	4000*		1660	8220	1.05%	4.2%

SMB data averaged from Clarke and coworkers (1975-85), Logothetis (1984)

Statistics and analytical methods

Ti (XRF) (Accurate to 30 ppm)

Zr (XRF) Precision $\pm 0.56\%$, accuracy -0.99% at 200 ppm; below 50 ppm, ± 10 ppm

Nb (XRF) (Quoted results rounded to 5 ppm intervals by XRAL)

Ta (INNA) Precision $\pm 10\%$

Y (INAA) Precision $\pm 0.79\%$, accuracy -0.44% at 50 ppm

F Results from New Ross-Vaughan Complex only.

and hydrothermal trends defined within the SMB provide an excellent regional framework for the geochemical discussions of Richardson, (1983) and Chatterjee and Strong (1984). The main objective of this contribution is to separate primary magmatic fractionation trends from hydrothermal or metasomatic (or both) alteration overprints in the Davis Lake Complex and to then evaluate the contribution of each of these processes to the greisen-hosted tin mineralization.

MAJOR ELEMENT INDICATORS

The differentiation index (D.I. of Thorton and Tuttle, 1960) increases with fractionation and is normally a useful trend of magmatic evolution within a cogenetic suite. The range of the Davis Lake Complex is 89.4-95.5. This is similar to that for monzogranite (86.8-92.7) and dykes and late intrusive phases (88.6-94.7) of the South Mountain Batholith, and higher than that of granodiorite (74.4-84.3) (Clarke and coworkers, 1975-1985).

Clarke and coworkers (1975-1985) have used Harker silica variation diagrams to illustrate magmatic fractionation trends in the SMB. However, within the Davis Lake Complex, other than TiO_2 , F_2O_5 and Al_2O_3 , the major element oxides show little variation (see

Fig. 8-11 as an example of this clustering) (Richardson, 1983). In the SMB, TiO_2 , Al_2O_3 and P_2O_5 decrease with increasing SiO_2 in granodiorite and monzogranite (Figs. 8-12, 13, 14) (Clarke and coworkers, 1975-1985). Several samples of the least altered Davis Lake biotite monzogranite lie at the extreme end of the magmatic trend defined by the South Mountain Batholith, with samples from other mineralized monzogranites (i.e. NRVC). However, the trends defined by the Davis Lake Complex reverse (Fig. 8-12 Al_2O_3) or intersect and are completely different from (Fig. 8-13, TiO_2 and Fig. 8-14, P_2O_5) the South Mountain Batholith. Instead, the Davis Lake Complex trends correlate well with processes that are coincident with increasing fluorine concentrations and pervasive albite-white mica alteration intensity observed macroscopically and petrographically.

The whole rock geochemistry of the mineral assemblages of both fresh biotite monzogranite and albite-white mica altered rocks are similar. Quartz, plagioclase, albite, K-feldspar and micas in the biotite monzogranite correspond to the quartz, albite, micas and topaz in the altered rocks. This suggests that the clustering of major element analyses is due to a simple local redistribution of Si, Ca, Al, K, Fe and Na from primary into alteration mineral phases. Similarly, mineralized, altered and fresh granites from the Erzgebirge tin district, Czechoslovakia show little separation on

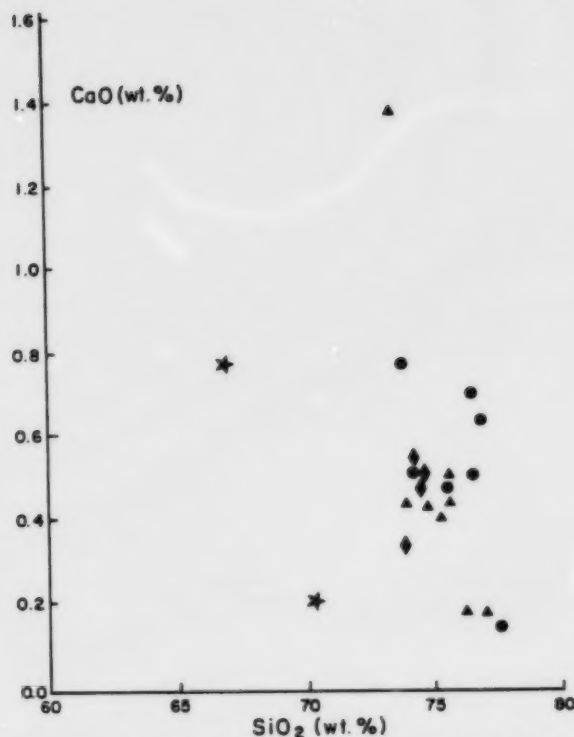


Figure 8-11. Harker silica variation diagram for CaO illustrating the clustering effect. Legend: Solid circles biotite monzogranite, Pointing up triangles regional blue leucocratic monzogranite, pointing down triangles blue leucocratic monzogranite below deposit, diamonds variably altered leucocratic monzogranite, stars-*greisen Davis Lake Complex.

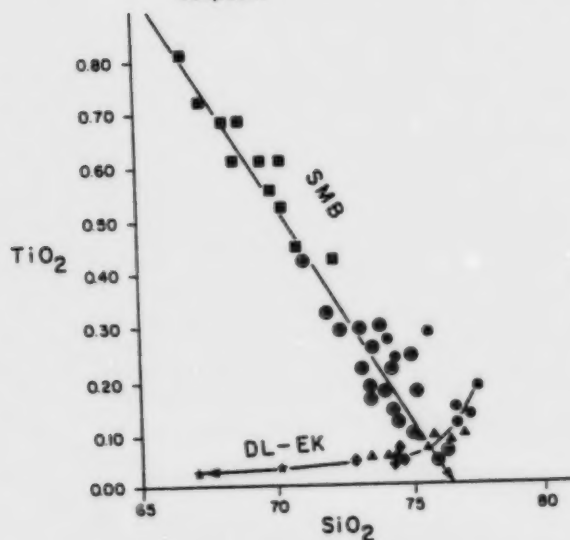


Figure 8-13. TiO_2 vs SiO_2 South Mountain Batholith and Davis Lake Complex. Legend as in Figure 8-12. Both components in weight percent.

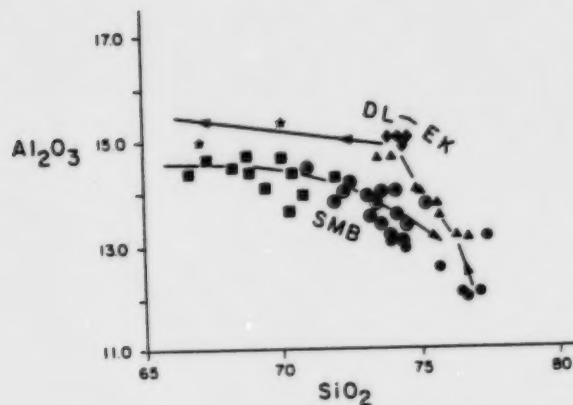


Figure 8-12. Al_2O_3 vs. SiO_2 , South Mountain Batholith and Davis Lake Complex. Legend as in Figure 8-11 open circles -undifferentiated monzogranite circles with enclosed stars- leucocratic monzogranite, squares- granodiorite: South Mountain Batholith.

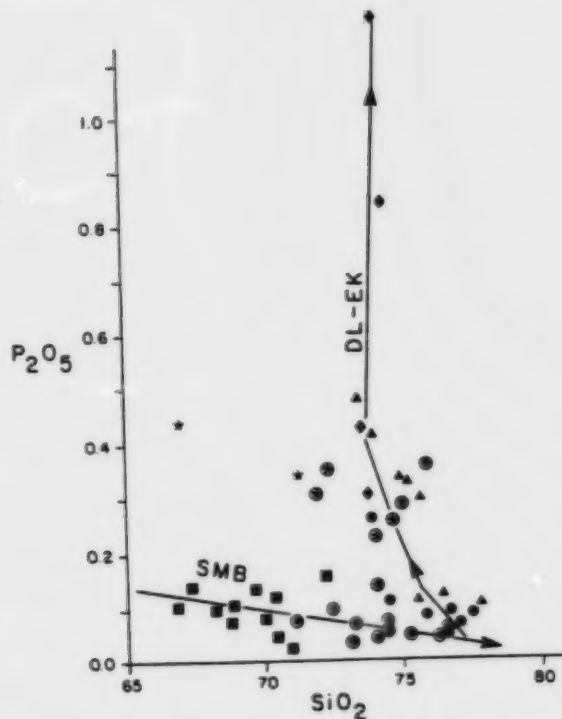


Figure 8-14. P_2O_5 vs SiO_2 , South Mountain Batholith and Davis Lake Complex. Legend as in Figure 8-12. Both components in weight percent.

normative Q-Ab-Or triangle diagrams until extremely altered (Stemprok, 1971).

Since phosphorus is an essentially non-substituting trace element, it is enriched during magmatic differentiation until a sufficient concentration is reached to precipitate phosphate minerals (commonly apatite). Watson and Capobianco (1981) indicate that felsic melts (70-80%SiO₂) reach apatite saturation at 0.14 with P₂O₅ at 750-900°C, independent of SiO₂ content. Saturation also correlates positively with temperature and perhaps CaO, and negatively with fluorine contents. Data from the SMB (Clarke and coworkers, 1975-1985) suggest that the initial parent magma was saturated early in the crystallization sequence (the granodiorite has 0.16 wt% P₂O₅) and precipitated apatite throughout its crystallization history. Biotite monzogranite of the Davis Lake Complex plots on the lower end of the magmatic trend that characterizes the South Mountain Batholith (Fig. 8-14). In the Davis Lake Complex and, to a lesser extent, the NRVC, phosphorus rises again to above magmatic saturation levels in leucomonzogranite and in variably altered leucomonzogranites, suggesting that it is being enriched by metasomatic or hydrothermal processes, or both. The altered leucocratic monzogranites of the Davis Lake Complex contain insufficient calcium to utilize all of this excess phosphate as fluorapatite. Instead, an unusual suite of phosphate minerals, such as triplite, vivianite, childrenite, and a new phosphate mineral, is present as well as fluorapatite.

In the Davis Lake Complex, alumina decreases with increasing silica content (Fig. 8-12). This element is an essential component of topaz, micas and the new phosphate mineral, suggesting that its unusual distribution is also controlled by metasomatic/hydrothermal processes.

TRACE ELEMENT INDICATORS

Trace elements such as Ti, Zr, Y, Nb and Ta are often used to document primary fractionation trends in other environments, since they are not generally transported by aqueous fluids or significantly affected by metasomatic alteration (Pearce and Norry, 1979). In the SMB, Ti and Zr decrease, (Clarke and coworkers, 1975-1985) and Ta increases (Muecke and Clarke, 1981) with magmatic fractionation. Ti, Zr and Ta data from the Davis Lake Complex are consistent with these trends (Richardson, 1983). Yttrium decreases in the Davis Lake Complex as would be expected and Nb increases with albite-white mica alteration, but decreases in greisen (Table 8-3) (Richardson, 1983). No comparative data are available for the SMB.

Boissavy-Vinau and Roger (1980) successfully used tantalum to indicate magmatic evolution in the albitized, stanniferous granites and pegmatites of the Marche area of the French Central Massif. There it ranges from 10 (least fractionated) to 10 000 (most fractionated) ppm.

In comparison, the SMB contains a maximum of 10 ppm (Muecke and Clarke, 1981) (Table 8-3). The Davis Lake Complex is also depleted in Ta, containing only 0.72-3.8 ppm in all alteration phases. At low levels such as these, the analytical uncertainty is high and the range is slight, so the potential usefulness of this element as a magmatic indicator is limited.

Titanium is similar in charge and ionic radius to Fe³⁺, Fe²⁺, Mn²⁺ and Sn⁴⁺. It easily substitutes into biotite, iron oxide minerals, cassiterite or micas. These minerals are common at East Kemptville as primary igneous or hydrothermal minerals, suggesting that significant Ti removal is unlikely even during alteration. Analyses by XRF are accurate to 30 ppm. The range in titanium throughout the SMB and Davis Lake Complex is from 4000-200ppm (Table 8-3). Numerous data are available.

The zirconium saturation level in felsic, peraluminous magmas is low (60 ppm), so zircons precipitate early in the crystallization sequence (Watson, 1979). Its behavior under metasomatic/hydrothermal conditions is unknown, but experimental and geological evidence indicates that F- and B-complexes are effective in mobilizing the geochemically similar HREE (Bandurkin, 1961; Mineevy, 1965; Alderton et al., 1980; Muecke and Clarke, 1981; Taylor and Fryer, 1983). In the Davis Lake Complex, zircon occurs as euhedral grains in biotite monzogranite. With albite-white mica alteration, it becomes rounded, then occurs as anhedral blebs. It is not found in strongly altered rocks within the deposit, suggesting that either the level of Zr was too low in these rocks to crystallize zircon or that it was removed by fluorine-rich fluids. The range of Zr in the SMB and the Davis Lake Complex is 200-20 ppm. Analytical accuracy is poor below 50 ppm (Table 8-3).

RARE EARTH ELEMENTS

The primary control to the REE distribution in highly evolved granitic rocks appears to be the presence of accessory minerals (Gromet and Silver, 1978; Fourcade and Allegre, 1981). Hydrothermal fluid-melt-rock interactions have a marked effect on the subsequent mobility and redistribution of the elements (Cornish granites: Alderton et al., 1980; South Mountain Batholith: Muecke and Clarke, 1981; Davis Lake-East Kemptville: Chatterjee and Strong, 1984; Richardson, 1983a, b, review; Taylor and Fryer, 1983). In the Davis Lake Complex, the REE distribution is characterized by a marked HREE depletion that can be correlated with accessory mineral habit, type of alteration and spatial location. The Eu/Eu* ratio remains relatively constant, suggesting that feldspar fractionation, normally an important magmatic process, was not important in this instance. Regional, least-to incipiently-altered biotite monzogranite has fairly flat REE patterns and a small Eu anomaly (Fig. 8-15, Table 8-4), similar to, but flatter than the average leucocratic monzogranite of the SMB (Muecke and Clarke, 1981). Zircon and apatite occur as unaltered

Table 8-4. Geochemical values for rare earth elements, Davis Lake Complex.

	BMG	BLM	VALM	GSN	
n=	6	9	6	Quartz	Topaz
La	17.6	2.9	2.65	3.9	5.7
Ce	49.1	13.5	10.7	8.6	13.1
Sm	12.8	4.2	2.9	2.3	3.9
Eu	0.6	0.3	0.4	0.04	0.3
Tb	0.9	0.4	0.2	0.0	0.5
Yb	4.0	0.17	0.1	0.1	0.2
Lu	0.7	0.01	0.03	bdl	bdl

bdl: below detection limit

euhedra in biotite from the Davis Lake Complex. Zircon will incorporate HREE preferentially over LREE, whereas apatite prefers the LREE (Hansen, 1978). Minerals that accept the LREE include monazite and magnetite (Mittlefehldt and Miller, 1983, Mahood + Hildreth, 1983).

In blue leucocratic monzogranite, biotite is replaced by white mica-chlorite-iron oxide aggregates that contain rounded zircon and apatite. Compared with the biotite monzogranite, these rocks are depleted in total REE content, especially HREE (Fig. 8-15). Sporadic small metamict halos in white mica aggregates are the only suggestion of previous zircon in variably altered leucocratic monzogranite (Richardson, 1983 a, b). Chatterjee and Strong (1984) report similar decreases in LREE, a slight decrease in the size of the europium anomaly and variations in HREE behavior.

The primary Davis Lake magma probably precipitated zircon and apatite early in the crystallization history until it was no longer saturated with zirconium and calcium. HREE were probably incorporated in zircon, LREE in apatite and LREE in magnetite or perhaps as yet unidentified monazite. With increasing fluorine contents (probably in an associated fluid phase) and progressive alteration, zircon was broken down, releasing Zr and associated HREE. Progressive alteration resulted in the continuous depletion of the HRCC. Muecke and Clarke (1981) used REE/fluid partitioning to explain similar, but not as marked REE depletion in the South Mountain Batholith. Fluorine contents of the Davis Lake biotite monzogranites are comparable to those of the biotite-muscovite monzogranite of the NRVC (Table 8-3-data from Logothetis, 1984). However, the blue leucocratic monzogranite of the Davis Lake Complex contains twice as much fluorine as leucomonzogranite from NRVC. Strongly altered leucocratic monzogranites from the deposit are even more fluorine-rich (Table 8-3). These higher fluorine contents could account for the more extreme depletion of HREE in the Davis Lake Complex.

East Kemptville greisen contains variable topaz, quartz and mica. Quartz-rich greisen is extremely REE depleted, perhaps because neither mica nor quartz can accept REE³⁺ ions. Topaz-

rich greisen seems to accept some REE, possibly substituting for Al³⁺ (Fig. 8-15). Rare fluorite occurs erratically in both types of greisen. The large difference in normalized europium may be due to Ca-Eu substitution in fluorite. However, overall, REE are apparently not concentrated in the minerals characteristic of East Kemptville greisen. Chatterjee and Strong (1984) report that East Kemptville greisen is low in total REE and depleted in HREE. It is difficult to ascertain which of their samples fit the strict definition of greisen used here.

At present, not enough data exist to fully evaluate the specific effects of fluid composition, albite-white mica and greisen alteration on the primary, accessory and alteration minerals in the Davis Lake Complex as suggested by Richardson (1983a) and Chatterjee and Strong (1984). Detailed work using mineral assemblages is necessary before further interpretation can be attempted. It is obvious though that the effects of magmatic fractionation and fluorine-rich hydrothermal fluids on accessory minerals have been important.

DISCUSSION AND CONCLUSIONS

Of the trace elements evaluated, titanium appears to be the best indicator of magmatic evolution and differentiation, based on the criteria defined previously. It decreases with fractionation as it crystallizes into mineral phases. It has only limited mobility as a result of alteration since it substitutes into iron oxide opaque minerals, alteration micas and cassiterite. It is readily determined by XRF and statistically valid with a wide range in the desired window (4000-200 ppm.).

In the Davis Lake Complex, the distribution trends for titanium appear to be linear. It is suggested that these trends reflect progressive magmatic differentiation from biotite monzogranite to the regional occurrences of blue leucocratic monzogranite to those marginal to and under the deposit to the variably altered leucocratic monzogranite to greisen. However, this trend is also concurrent with increasingly intense, pervasive albite-white mica alteration and rising F and P₂O₅ contents (Figs. 8-14, 16). This correspondence of magmatic fractionation,

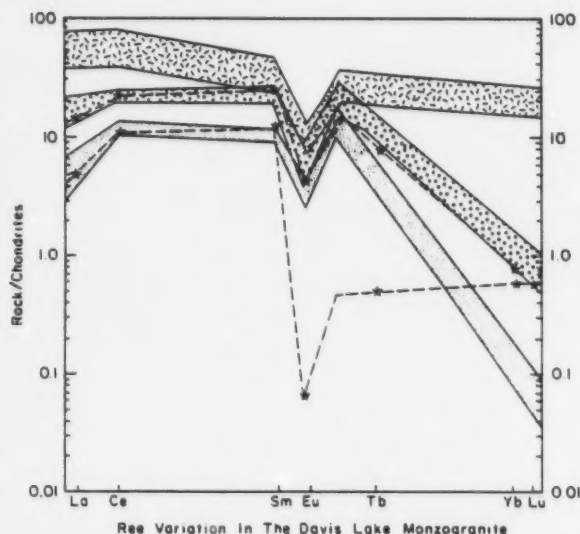


Figure 8-15. Composite profile of REE distribution in Davis Lake Complex. Top cluster-6 biotite monzogranites, Middle cluster-4 blue leucocratic monzogranites, Lower cluster-4 variably altered leucocratic monzogranites. Uppermost individual pattern: topaz greisen, lowermost individual pattern: quartz greisen.

intensity of alteration and spatial location suggests that there is not a clear-cut temporal separation of magmatic and hydrothermal processes in the Davis Lake Complex. Rather, the crystallization of the biotite monzogranite was dominated by magmatic processes and is similar to monzogranites of the NRVC. At the other end, cassiterite-mineralized greisen within the deposit is a hydrothermal, subsolidus end product chemically derived from the variably altered leucocratic monzogranite in the immediate vicinity.

In situ fractionation of highly evolved granitic melts such as the Davis Lake monzogranite commonly results in the concentration of incompatible elements in hydrothermal fluids. The degree of separation of these phases from crystal or cumulate phases appears to be a critical factor controlling the development of mineralization in granitic rocks (McCarthy and Hasty, 1976; Groves and McCarthy, 1978; Hildreth, 1981). In the Davis Lake Complex, the excellent correlation among the spatial location, degree of fractionation, intensity of albite-white mica alteration and greisen, and degree of mineralization suggests that a progressive process of interacting, on-going fractionation, crystallization, fluid phase separation and collection, and subsequent alteration played an important role in the genesis of the East Kemptville deposit.

Throughout the Davis Lake Complex, fluids collected under several local inflections in the contact (e.g. Dog Lake, Big Meadow Brook),

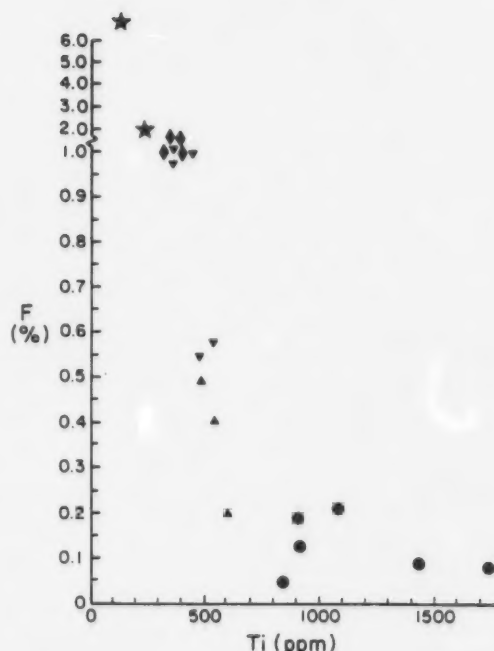


Figure 8-16. F vs Ti, Davis Lake Complex. Ti in ppm., F in %. Legend as in Fig. 8-11.

resulting in small zones of albite-white mica alteration and greisen. However, only within the area of the East Kemptville deposit were chemical and physical conditions optimal for the more advanced fractionation, crystallization and continued in situ fluid evolution necessary for more intense albite-white mica alteration, greisen and economic tin mineralization. The chemically unreactive, impermeable quartzite of the Meguma Group acted to cap this. Underneath the contact, an unusually fluorine-phosphate-metal rich-fluid accumulated, possibly through liquid fractionation by immiscibility, vapor/fluid separation, or thermogravitational diffusion and the Soret effect, aided by convective overturn (Hildreth, 1981) or through fractional crystallization (Groves and McCarthy, 1978). It altered the crystalline roof zone, producing zones of massive greisen immediately below the granite-metasediment contact. Eventually, the pore-fluid pressure exceeded that of the combined lithostatic pressure and the tensile strength of this crystalline framework and fracturing occurred. Field observations indicate that fracturing was largely restricted to the granitoid rocks and that the metasedimentary rocks were not extensively brecciated. The fluid migrated down from the contact zone in oriented fracture sets, greisenizing the monzogranite marginal to the fractures and overprinting previous albite-white mica alteration.

Although the intrusive relationships are not positively established, the Davis Lake Complex does appear to be comagmatic with the SM8,

but more fractionated and specialized. The primary geochemical enrichment and metalliferous nature of the parent Davis Lake magma, extreme fractional crystallization, early fluid phase separation and continual fluid evolution resulted in geochemically distinct, but spatially and temporally, related, albite-white mica alteration that was locally overprinted by fluorine-metal-rich greisen. The geometric shape of the granite-metasediment contact and the impermeable and chemically unreactive nature of the overlying quartzites were also important in localizing mineralization at East Kemptville.

ACKNOWLEDGMENTS

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CHAPTER 9

PETROLOGY OF THE POLYMETALLIC QUARTZ-TOPAZ
GREISEN AT EAST KEMPTVILLE

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INTRODUCTION

A large (58 million tonnes) polymetallic quartz-topaz-cassiterite greisen deposit at East Kemptville (44° 16'N; 65° 41'W) in Yarmouth County of Nova Scotia was discovered by Shell Canada Resources Limited in 1978. Rio Algom confirmed the possibility of economic recovery of tin from the ore, thus establishing the first North American primary tin deposit. The tin production from East Kemptville, as projected, would contribute approximately 8.3 per cent of the global bedrock tin production, thus ranking third behind Bolivia and Australia.

Despite the economic significance of the greisen body (1.5 km x 0.5 km x 0.1 km) it is important to study the deposit in its regional setting together with various processes (cf. Clarke and Chatterjee, 1985), which evolved through the magmatic and greisenization stages to give rise to the East Kemptville polymetallic tin deposit. The present contribution deals with the regional and local geology, distribution, form and composition of gangue and ore minerals, and geochemical characteristics of the granitoid rocks surrounding the East Kemptville tin deposit. Other details on petrology, geochemistry, REE mobility and complexing, and isotope geochemistry have been elaborated by Chatterjee (1980), Richardson et al. (1982), Muecke and Chatterjee (1983), Chatterjee and Strong (1984a), Richardson (1985) and Chatterjee and Strong (1985a).

GEOLOGICAL SETTING

The East Kemptville tin deposit is located within a well defined, northeast trending zone which measures approximately 70 km long and 8 to 10 km wide (Chatterjee, 1983). The zone is dominated by Cambro-Ordovician metasedimentary rocks of the Meguma Group and stratigraphic relationship suggests that it is part of a prograding turbidite fan progressing upwards from a distal to a suprafan environment (Walker, 1979). The earliest structure observed in these rocks is a bedding - parallel foliation produced by loading. However, the metasedimentary rocks lie on the western limb of a major NNE-SSW trending anticline. The minor folds are generally tight, upright, north-south trending with gently plunging fold axes. Regional low pressure metamorphism (415 - 400 Ma) followed the folding producing quartz-muscovite-biotite-andalusite $\pm$ garnet $\pm$ cordierite $\pm$ staurolite assemblages in lithological units of appropriate composition (Reynolds and Muecke, 1978). Intrusive into the regionally metamorphosed sediments are the granitoid rocks of the South Mountain Batholith (SMB) which has been dated at 372 ± 2 Ma - 361 ± 2 Ma (Clarke and Halliday, 1980; Reynolds et al. 1981). The SMB is represented by the plutonic rocks of the Davis Lake Complex at the northern end of the tin domain and the 313 ± 3 Ma old Wedgeport Pluton (Keppie et al. 1983) marks the southern termination. For details on the geology of southwestern Nova Scotia, the readers are referred to the work by Taylor (1967), and Taylor and Schiller (1966).

Within the tin domain of southwestern Nova Scotia, polymetallic tin mineralization is physically controlled by joints, shear zones, and the contact between the granitic and the meta-sedimentary rocks. The mineralization is either intrabatholithic (endo-contact) or peribatholithic (exo-contact).

In the following pages the geology of the Davis Lake plutonic Complex with which the polymetallic quartz-topaz greisen at East Kemptville is spatially associated, is elaborated and the geology of the Wedgeport Pluton is discussed in Chapter 11.

The geological setting of the East Kemptville tin deposit and its relationship with the granitic rocks of the SMB within a radius of approximately four miles is shown in Figure 9-1. Within this limited area, the following sequence is observed: coarse grained megacrystic biotite granodiorite, medium- to coarse-grained biotite monzogranite, leucocratic muscovite-biotite-topaz monzogranite, and quartz-topaz greisen. Though this evolution is not established on the basis of geochronology, the petrological evolution as discussed below is almost identical to that established for the SMB (McKenzie and Clarke, 1975; Clarke and Halliday, 1980). The topaz-bearing monzogranite shows pronounced effects of hydrothermal alteration, and is represented petrographically by porphyritic biotite granite, argillized and sericitized granite, silicified granite, albitized granite, and development of the polymetallic tin-greisen. Airborne gamma ray response in the area is marked and illustrates many features which conforms to the evolution of the granitoid rocks mentioned above. In Figure 9-1 we have plotted the eU/eTh ratio contours (simplified from Geological Survey of Canada Open File Map #35121 (3) G, 1982) which show that the granodiorite is the least evolved rock (eU/Th = 0.2) followed by biotite monzogranite (eU/Th = 1.0), and the topaz-bearing monzogranite in which the ratio generally less than 1, and finally the greisen zone itself which is marked by a very pronounced anomaly and the ratio reaches a high of 2.4. This systematic change within a relatively small area is significant, as will be discussed later, and is essentially controlled by the divergence of the geochemical behaviours of uranium and thorium in the granitic melts and greisen fluids (Chatterjee and Muecke, 1982; Muecke and Chatterjee, 1983; Chatterjee and Strong, 1984b; Ford and O'Reilly, 1985).

PETROGRAPHIC DESCRIPTIONS

Biotite Granodiorite

This rock type is the most widespread and volumetrically the most abundant unit in the SMB. McKenzie and Clarke (1975), Clarke and Halliday (1981) and Chatterjee et al. (1982) have described this unit in detail and the reader is referred to the above publications for a more comprehensive treatment (mineralogy, major and trace element characteristics and isotope geochemistry). The granodiorite of the Davis Lake Complex consists of quartz, andesine, K-feldspar, and biotite as major constituents. Accessory minerals are apatite, zircon, pyrite and chalcopyrite. Textures range from xenomorphic to

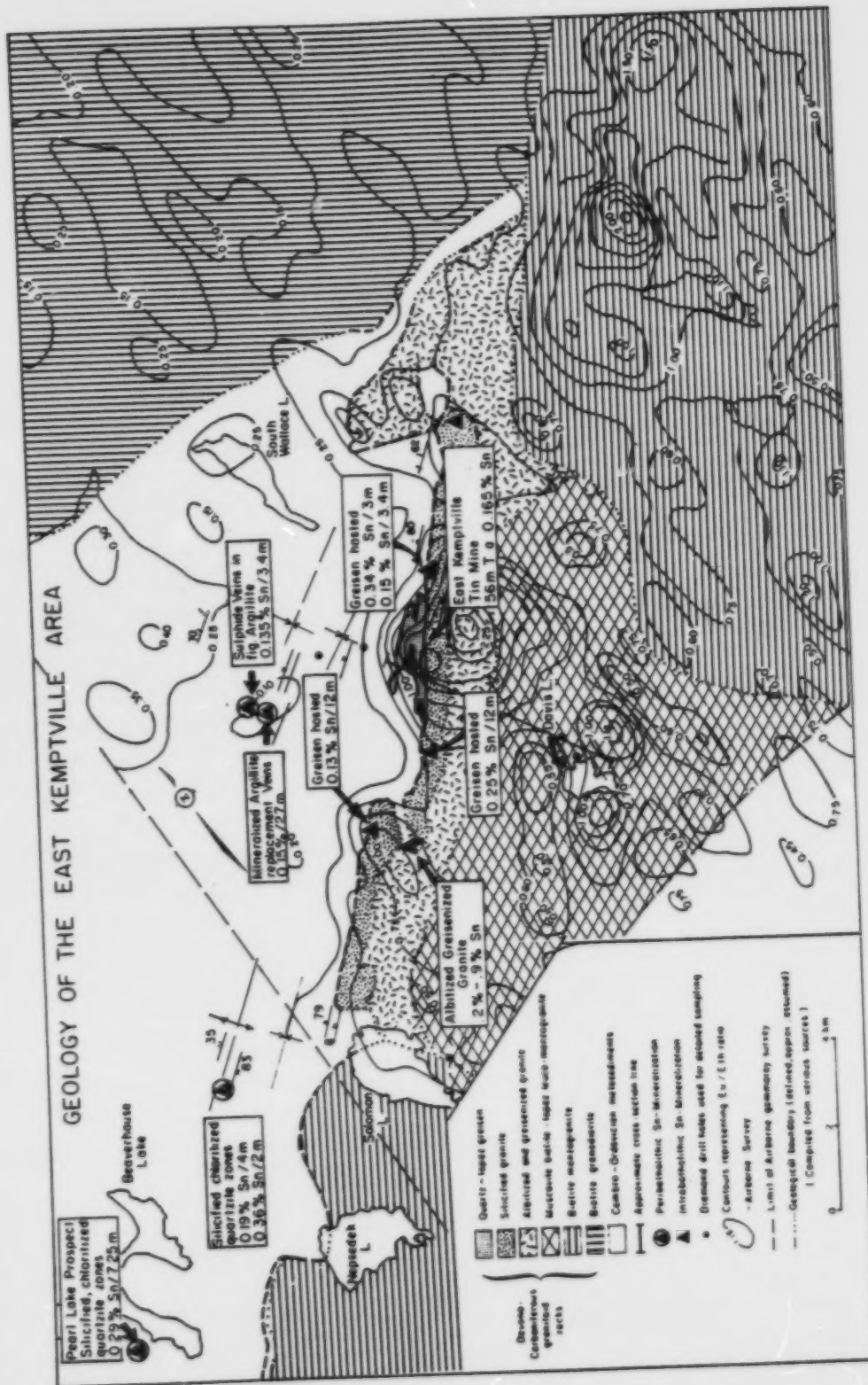


Figure 9-1. Geological map of the East Kemptville area.

hypidiomorphic, and both oscillatory and patchy zoning are common in the plagioclase. Myrmekitic intergrowths are present adjacent to potassium feldspar.

Biotite Monzogranite

The leucocratic biotite monzogranite consists of 40-45% quartz, 30-35% K-feldspar, 20-25% oligoclase and less than 3% biotite, muscovite, apatite and zircon. Small amounts of sericite, pyrite and chalcopyrite are accessory minerals. Quartz either forms irregular grains that replace oligoclase, or it occurs interstitially. Most of the quartz is strained. Irregular grains of K-feldspar are as much as four centimetres in diameter and replace both oligoclase and quartz. Oligoclase occurs as subhedral to euhedral crystals having an average composition of Ab76. Oscillatory and patchy zoning show in most crystals, and some have rims of normally zoned material grading into albite on the outer edge. Irregular shreds of biotite as much as six millimetres long are pleochroic from light yellowish-brown to dark brown. Zircon and apatite occur only as minute grains in with pleochroic halos. Pyrite and chalcopyrite occur as fine dust along grain borders, and in altered cores of oligoclase or as larger grains along fractures. The cores of the plagioclase are often altered to sericite and calcite.

MUSCOVITE-BIOTITE-TOPAZ MONZOGANITE

This rock type shows marked compositional and textural variations and these are discussed in the following discussion. The rock types constitute the paraintrusive suite of the Davis Lake Complex.

Biotite Granite

The most abundant rock type is porphyritic biotite granite. It is generally pale grey to dark grey in colour and massive to weakly foliated. Most of the conspicuous phenocrysts of the rock are K-feldspar, but locally quartz also forms phenocrysts and the rock has 'quartz-porphyry' aspect. The rock is composed generally of equal proportions of quartz, plagioclase and perthitic K-feldspar which together constitute approximately 95 per cent of the rock. Muscovite, biotite, topaz, fluorite, andalusite, pyrite, zircon and rutile in approximate decreasing order of abundance, comprise the accessories.

The K-feldspar has prominent cross-hatched twinning and, on the basis of X-ray diffraction data, plots toward the orthoclase end of the orthoclase-maxima microcline structural composition series of Wright (1968). On the average the K-feldspar contains approximately 18 per cent albite as determined by microprobe analysis. Most of the plagioclase occur in fine- to medium-grained subhedral to euhedral crystals and approximately 20 per cent of the crystals show deformation of the twin lamellae (Figure 9-2A,B). The composition of plagioclase as determined by microprobe analysis ranges from Ab76 to Ab84. The discrete plagioclase crystals are not zoned but an increase in Or component from the core to

the inner rim and then a decrease in the outer rim, in some grains, indicates that they are not uniform in composition, although the overall composition is approximately calcic oligoclase. Generally, biotite and muscovite are equally abundant and interstitial, but some occur in plagioclase crystals where it appears to be secondary and when plotted on a Al_2O_3 -FeO-MgO diagram they all fall in the BM field of Nocholds (Chatterjee and Strong, 1984a). Muscovite occurs as irregular masses cross-cutting biotite or K-feldspar, and sparse sericite occurs as a separate phase or in aggregate with topaz and andalusite (Figure 9-2C, D). Andalusite and topaz occur in close association and share fractures (Clarke et al. 1976) which is suggestive of their primary magmatic minerals. From the above description it is fair to conclude that most of the texture of the leucocratic topaz-bearing monzogranite is due to magmatic crystallization.

Argillized and Sericitized Granite

This rock type is nearly massive, homogeneous granular muscovite-biotite granite. In this section, the texture of the rock is hypidiomorphic against discrete plagioclase grains. The potassium feldspar phase of the perthite is pale pink and dust, and the plagioclase lamellae are almost white and clear. The K-feldspar contains small tabular crystals of plagioclase, some of which are cut by the albite lamellae of the perthite. Locally, either the K-feldspar or albite phases of the perthite apparently replace discrete sodic plagioclase crystals.

The most abundant mineral is medium grained microcline microperthite of vein or patch type that occurs in subhedral grains and is xenomorphic against discrete plagioclase grains. The potassium feldspar phase of the perthite is pale pink and dust, and the plagioclase lamellae are almost white and clear. The K-feldspar contains small tabular crystals of plagioclase, some of which are cut by the albite lamellae of the perthite. Locally, either the K-feldspar or albite phases of the perthite apparently replace discrete sodic plagioclase crystals.

Plagioclase, exclusive of that found in perthite intergrowth, forms almost white, subhedral to euhedral crystals which are polysynthetically twinned, generally by the albite law. The plagioclase is slightly coarser grained than the perthitic microcline, but also has a wider range in grain size. The albite content of the plagioclase ranges from Ab89 to Ab97 and the orthoclase component ranges from 0.6 to 8.1 per cent. As can be seen from Figure 9-3, the plagioclase in the altered granite is albite with relatively high 'Or' content and is distinctly different in composition than the biotite granite.

Quartz generally occurs as subhedral to euhedral grains or aggregates and, rarely, it is idiomorphic against potassium feldspar. Quartz also forms symplectic border zones around some biotite crystals and locally has lobate convex contacts against plagioclase, potassium feldspar and biotite. Muscovite typically forms euhedral to subhedral flakes, as thin rims on biotite, in thin zones cutting across biotite flakes or in



Figure 9-2. Photomicrographs showing the textural relationships in plagioclase feldspar. Note the deformation of twin lamellae, development of myrmekite and replacement of biotite by muscovite.

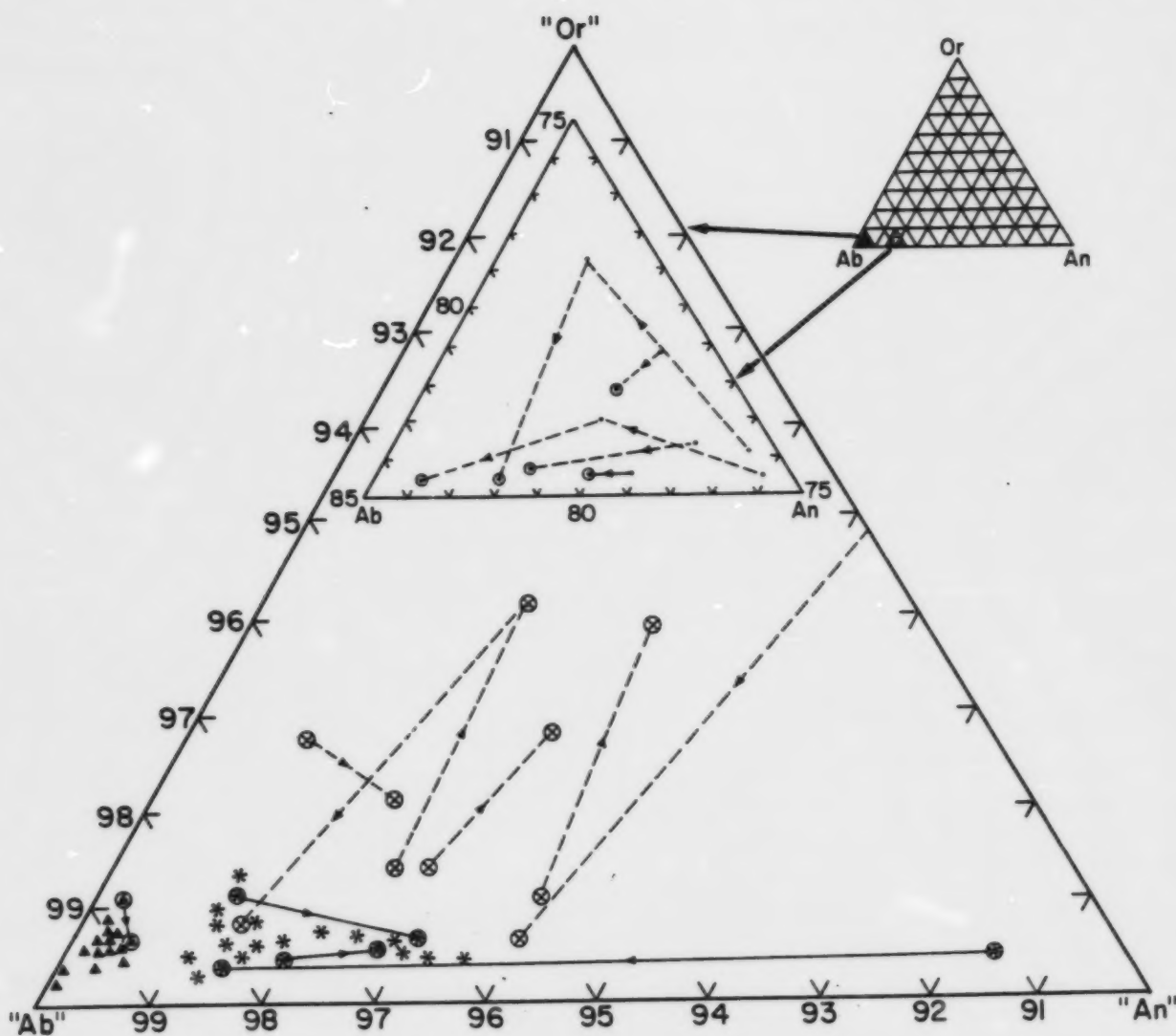


Figure 9-3. Ternary Or-Ab-An diagram showing the composition of plagioclase feldspar from the granitoid rocks of the Davis Lake area. (For explanation see the text).

crystals intergrown with biotite. It is accompanied by a group of minerals which are mainly interstitial or occur with interstitial minerals, namely fluorite, apatite, topaz and cassiterite. Small crystals of fluorite, white mica and topaz also occur in albite (Figure 9-11), particularly along the twin lamellae and in this mode they are considered to be infiltrational.

Silicified Granite

The silicified granite is fine- to medium-grained, xenomorphic (except for feldspar) and equigranular. It appears to be associated with the hydrothermal introduction of quartz-topaz-cassiterite sulphide veins (Figure 9-4A, B). In this type, thin quartz veinlets are common, and rock are discontinuously sheeted and interfingering in places. In thin section, feldspar is partly clouded by clay minerals and contains scattered grains of muscovite. Plagioclase (Ab₉₀ to Ab₉₅) forms short subhedral grains that are untwinned. K-feldspar is mainly micropertitic, has undulatory extinction, and the grain boundaries range from orthogonal to very irregular and interstitial. Biotite occurs as remnant grains that have been partially replaced by muscovite and chlorite (Figure 9-5). Accessory minerals consist of fluorite, topaz, cassiterite, chalcopyrite, pyrite and pyrrhotite.

Albitized Granite and Albitites

In hand specimen, the typical albitized granite is pale grey, massive, saccharoidal textured and has no visible mafic minerals (Figure 9-4C, D), and occasionally some evidence of cataclasis is seen. Microscopically, the texture is hypidiomorphic granular and micrographic, and both may be present in the same thin section. In the dominantly hypidiomorphic rock, plagioclase is the most idiomorphic constituent, forming coarse grained euhedral to subhedral tabular grains. Potassium feldspar has a hyalophitic relation with the plagioclase and is in subhedral to euhedral grains. Plagioclase feldspar and quartz both locally appear to replace the potassium feldspar. Two types of micrographic textures are common. Type I has a irregular vermicular form which probably resulted from extensive replacement of tabular sodium plagioclase by quartz, and the original tabular outlines of the plagioclase can be inferred from the optical continuity of the plagioclase remnants. Type II consists of quartz and clear, untwinned albite intergrowths in a radial or rosette-like form, which in places encloses a nucleus of older plagioclase or potassium feldspar. The albite is twinned on the Carlsbad, Albite-Carlsbad and Albite-Ala B laws. Its average Ab content is Ab₉₉ and the 'Or' content is vanishingly small.

The presence of albitites and albite breccia is rather unexpected with the granitoid rocks of the SMB because magmatic albitites are generally associated with the alkalic carbonatite complexes. However, the postorogenic fold belt type tin districts in Europe and the USSR are all characterized by albitized granites or albitites, or both (Taylor, 1979). Taylor (1979) aptly pointed out that examples of albitites outside of Europe and the USSR are rare, although this

probably reflects the problems of recognition rather than an absence. The phenomenon has recently been noted at several points within the eastern Australian tin provinces, after seven years of active exploration since the initial discovery of tin mineralization.

The albitites are leucocratic, coarse grained, porphyritic, albite-quartz-muscovite rocks. They are characterized by complete absence of K-feldspar. In the porphyritic variety crystals of albite, those up to three centimetres in size or crystal clusters are set in a matrix of lath-like albites and quartz. Small amounts of interstitial quartz, biotite, muscovite, phenakite, bertrandite, cassiterite, apatite, fluorite and cryolite are often present. The albitites grade into albitite megabreccia. Mineralogically the breccia is similar to albitite except that quartz is more abundant.

Quartz-Topaz Greisen

The polymetallic quartz-topaz greisen veins are found in an area that covers approximately 15 km². Despite this broad distribution, the only cluster measuring 1500 m x 500 meters (i.e. the East Kemptville tin deposit) occurs beneath the northwestern contact of the Davis Lake Complex with the metagreywacke of the Cambro-Ordovician Meguma Group. The greisen veins (<1 cm to <1m) have been intersected in drillholes to a depth of 150 metres.

Individual greisen veins within the East Kemptville cluster are narrow (no more than 50 cm) and mostly tabular. The veins pinch and swell locally and may vary greatly in thickness in a short distance. The greisen veins occur singularly in most places within the cluster, although in a few places they appear to occur as short an echelon segments. Where the veins are locally close together, the granite has a banded appearance. These and other morphological details of the greisen veins are given by Richardson et al. (1982). The relationship of the veins in the greisens is almost identical to the structure of greisens characteristic of simple and telescoped zonations (Shcherba, 1970, p.133).

The quartz-topaz-cassiterite veins are generally bounded by a selvage of hydrothermally altered rocks which became conspicuous when the samples are stained with cobaltinitrate. The granitic rocks adjacent to the veins have been strongly altered so that virtually all magmatic minerals have been replaced by hydrothermal quartz, topaz, muscovite, chlorite, fluorite and tourmaline.

Minerals of the polymetallic quartz-topaz greisen at East Kemptville comprise three groups: primary gangue minerals (quartz, topaz, muscovite, Ca-Mg carbonates, complex phosphate minerals, fluorite, tourmaline, chlorite and albite), primary ore minerals (cassiterite, chalcopyrite, sphalerite, wolframite, tetrahydroite-tennantite, stannite and galena. These are generally associated with other sulphide minerals, viz. pyrite, arsenopyrite, pyrrhotite, molybdenite), and secondary minerals (illite,



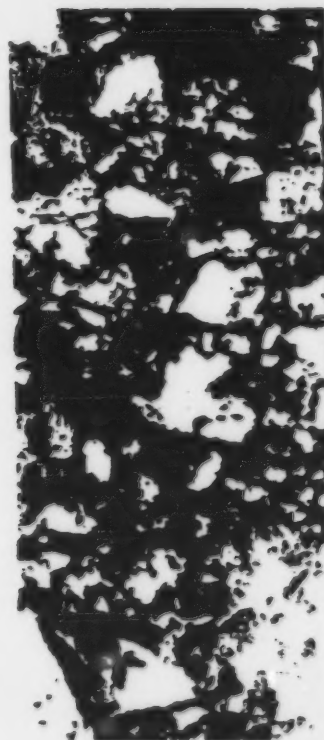
A



B



C



D

Figure 9-4. Photographs illustrating the silicification and development of quartz-topaz-cassiterite veins. The photographs also illustrate the albitization of rocks and brecciation of the albitite.

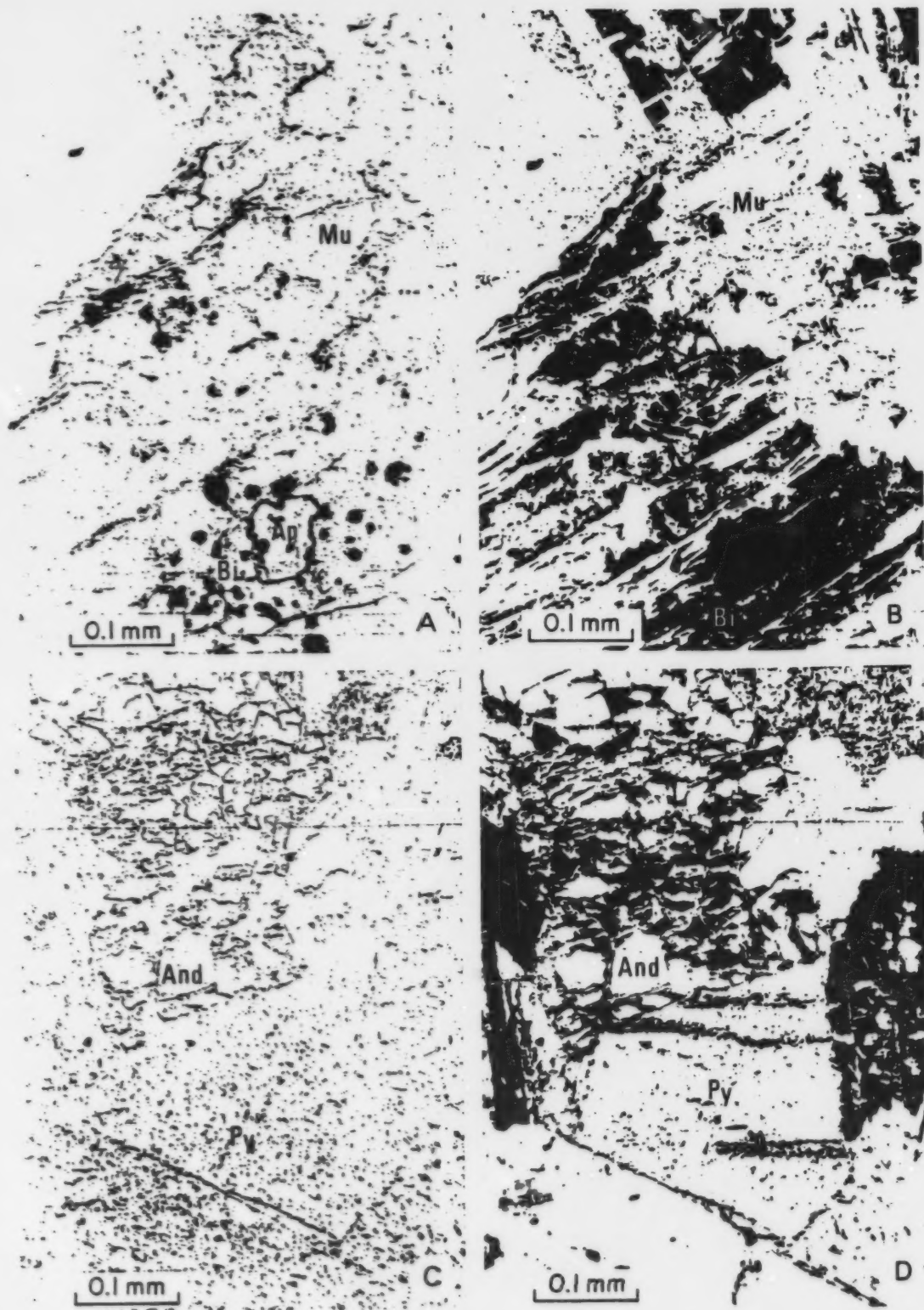


Figure 9-5. Photomicrographs illustrating the replacement of biotite by muscovite. Note the relict grains of apatite and zircon surrounded by pleochroic haloes. The figure also illustrates the alteration of andalusite to pyrophyllite.

dicite, stilbite, rhodochrosite, and pyromorphite). Also within the primary ore minerals, inclusions of native bismuth, argentite, wolframite, muscovite, quartz, topaz, tourmaline are generally present. The paragenetic sequence of the minerals is illustrated in Richardson (1985).

MINERAL CHEMISTRY

The composition of selected minerals in 29 representative rock samples were determined by electron microprobe analysis using a Cambridge Microscan 5 electron microprobe employing wavelength dispersive and energy dispersive systems and the EDATA 1 data reduction program. Corrections for atomic number, absorption and fluorescence were applied during data reduction. In order to calculate the mineral formulas, all iron was considered as ferrous and, for the phyllosilicates minerals cations, proportions were computed on an anhydrous basis assuring the ideal proportion of structurally-bound water.

PLAGIOCLASE

Representative analyses of plagioclase composition from the Davis Lake Complex are given in Table 9-1 and plotted in Figure 9-3. Oxide totals range from 97.53 to 101.89, cation totals (based on 32 oxygen) range from 19.768 to 20.047 and the average 19.986 is in excellent agreement with the ideal value of 20.0.

As to be expected the most calcic plagioclase observed = An₃₅ is from the biotite-granodiorite and compositional core to rim zoning represents a range from An₃₅ to An₂₁. This range in composition and zoning of the plagioclase is similar to the composition of plagioclase reported for the unaltered biotite granodiorite of the Millet Brook area. The average composition of the plagioclase from the biotite monzogranite is An₂₄ and it shows normal core to rim zoning. The composition of plagioclase from the muscovite-biotite-topaz monzogranite and the quartz-topaz greisen is plotted in Figure 9-3. It is clear that the composition of plagioclase in various rock types is distinctly different.

BIOTITE

Biotite occurs as euhedral to subhedral crystals in biotite granodiorite and monzogranite, and frequently contain inclusions of apatite, zircon and quartz. In the leucomonzogranite biotite coexists with muscovite, and the large flakes are generally devoid of inclusions but infiltrated by quartz, topaz, fluorite, cassiterite and various sulphide minerals. In the greisen, biotite is replaced by muscovite and chlorite, and in many cases the original grains are preserved together with their original inclusions. Figure 9-5A, B illustrates clearly that biotite has altered along cleavage planes to muscovite and perhaps this is in agreement with the interaction of the hydrothermal fluids with the previously crystalline rocks.

Detailed discussions on biotite composition from the East Kemptville deposit was presented by Chatterjee and Strong (1984a), and Strong and Chatterjee (1985). Representative microprobe

analyses and structural formulae are given in Table 9-1 and plotted in Figure 9-6. The systematic variation in biotite composition from the Davis Lake Complex is considered in a Ti vs. F/FM variation diagram. Other characteristics such as high Al content which is typical of biotite that coexist with other aluminosilicates (a feature well documented for the SMB), is obvious in the data.

The Ti ranges from 0.08 to 0.59 and the Fe/Fe+Mg ranges from 0.64 to 0.92 for the Davis Lake Complex. This Ti vs. F/FM diagram illustrates the nature of Ti⁴⁺ substitution in biotite for (Fe,Mg)²⁺ and also this variation probably follows the bulk composition of the rocks. The Ti⁴⁺ substitution in biotite generally follow one of the following reactions: $Ti^{IV} + 2Al^{IV} = Mg^{VI} + 2Si^{IV}$, $Ti^{IV} + [] = 2Mg^{VI}$ and $Ti^{IV} + 2O^{2-} = (Fe,Mg)^{VI} + 2OH^{-}$. Also there does not appear to be any correlation between Ti and Si^{IV} and Fe, or total octahedral cations, a relationship which should reflect coupling of Ti substitution with a vacancy. Thus, we have to conclude that probably TiO₂ substituted for (Fe, Mg)(OH₂).

MUSCOVITE

Muscovite is a common component in all the rock types except biotite granodiorite of the Davis Lake Complex. It is generally colourless but pale silvery or pale yellowish green varieties also occur. In many quartz-topaz greisen it forms a thin selvage (<1 cm to 1 cm thick) on the vein walls. Often the flakes grow normal to the vein margins but the c-axis of the crystals are roughly parallel to the margin. In places, the muscovite in the greisen veins show plumose habit. Textural relationships illustrated in Figures 9-6C, D and 9-7 and discussed in the petrographic section would suggest that most of the muscovite is secondary.

Ideally, muscovite is a hydrous potassium-aluminum silicate devoid of octahedral divalent cations. However, many investigations (e.g. Turner, 1968; Nash, 1973; Pinsent and Smith, 1975; Miller et al. 1981; Clarke et al. 1976; Clarke, 1981) have indicated that not only do the potassic dioctahedral micas depart significantly from the theoretical muscovite composition, but most muscovite in peraluminous granite product of is a subsolidus breakdown of aluminosilicates, e.g. andalusite, topaz (Clarke et al. 1976). Miller et al. (1981) reviewed compositional variations in muscovite from 16 different plutons, and suggested that magmatic muscovites are richer than secondary hydrothermal muscovites in Ti, Na and Al, and poorer in Mg and Si. Strong and Chatterjee (1985) showed that while most of the muscovites, associated with mineralized granitoids, plot within the secondary Mg-Ti-Na field outlined by Miller et al. (1981), each deposit has its own separate compositional range (Fig. 9-8).

The departure from the theoretical muscovite composition is essentially due to the isomorphous entry of the celadonite component, which results in an excess of Si, a depletion of Al, and entry of Mg and Fe into the muscovite lattice, e.g.

162 Granites & Mineral Deposits SW N.S.

Major Oxides (Weight per cent)								
	1	2	3	4	5	6	7	8
SiO ₂	38.64	38.61	38.29	37.95	36.01	36.69	36.58	36.40
TiO ₂	1.53	1.39	1.34	1.51	0.73	0.69	1.42	0.80
Al ₂ O ₃	15.88	15.98	16.49	16.38	18.39	17.93	19.98	19.53
FeO	21.36	21.30	20.94	21.49	29.02	29.16	27.95	28.16
MnO	-	0.06	-	-	0.36	0.34	0.60	0.43
MgO	10.33	9.76	9.26	9.38	1.86	2.06	1.41	1.84
CaO	-	-	-	-	0.07	0.09	-	-
K ₂ O	9.80	10.03	10.02	9.88	8.99	8.75	9.43	9.23
H ₂ O	4.01	3.99	3.96	3.96	3.78	3.80	3.88	3.84
Total	101.55	101.12	100.30	100.55	99.21	99.51	101.25	100.23
Structural Formulas								
Si	5.776	8.000	5.803	8.000	5.791	5.742	8.000	5.708
Al	2.224	2.197	2.209	8.000	2.258	2.292	8.000	2.216
Al	0.573	0.633	0.730	0.663	1.143	1.115	1.274	1.265
Ti	0.172	0.157	0.152	0.172	0.087	0.082	0.165	0.094
Fe	2.670	2.677	5.661	2.649	2.719	3.847	3.845	3.605
Mn	-	0.008	-	-	0.048	0.045	0.078	0.057
Mg	2.302	5.717	2.186	2.087	5.618	2.116	5.670	0.439
Ca	-	-	-	-	0.012	0.015	0.324	5.447
K	1.868	1.868	1.923	1.923	1.933	1.907	1.907	1.818
H	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
O	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000
Components								
Phlo	46.293	44.884	44.007	43.755	10.138	11.067	8.088	10.288
Ann	53.707	54.959	55.943	56.245	88.747	87.895	89.956	88.345
Mn	-	0.157	-	-	1.115	1.038	1.956	1.366
Major Oxides (Weight per cent)								
	9	10	11	12	13	14	15	16
SiO ₂	37.95	37.58	38.14	36.87	29.51	27.22	27.71	27.51
TiO ₂	0.65	0.34	-	-	-	-	-	-
Al ₂ O ₃	18.72	21.99	22.48	30.90	19.72	19.74	18.66	18.59
FeO	27.14	24.70	25.02	19.90	38.00	40.48	39.78	39.45
MnO	0.48	0.62	0.58	0.23	0.88	0.65	0.81	0.11
MgO	2.60	1.31	1.31	0.79	2.51	2.38	2.16	4.32
CaO	-	-	-	-	-	-	-	0.10
K ₂ O	10.08	9.32	9.44	-	4.20	1.75	3.21	-
H ₂ O	3.90	3.91	3.96	3.99	3.62	3.50	3.48	3.47
Total	101.52	99.77	100.93	92.68	98.44	95.72	95.81	93.55
Structural Formulas								
Si	5.823	5.756	5.770	5.541	4.887	4.653	4.776	4.746
Al	2.177	8.000	2.244	8.000	2.459	8.000	3.337	8.000
Al	1.207	1.175	1.778	3.013	0.736	0.649	0.566	0.525
Ti	0.075	0.039	-	-	-	-	-	-
Fe	3.482	3.184	3.185	2.501	5.283	5.800	5.734	5.692
Mn	0.082	0.080	0.074	0.029	0.123	0.094	0.118	0.016
Mg	0.595	5.421	0.299	5.307	0.295	5.313	0.177	5.720
Ca	-	-	-	-	0.620	0.742	0.608	7.151
K	1.973	1.973	1.821	1.821	1.822	1.822	-	0.887
H	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
O	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000
Components								
Phlo	14.366	8.440	8.356	6.536	10.316	9.347	8.661	16.291
Ann	84.129	89.290	89.542	92.382	87.629	89.302	89.494	83.473
Mn	1.507	2.270	2.102	1.081	2.055	1.451	1.846	0.236

Table 9-1. Representative microprobe analyses of biotite from various rock types from the Davis Lake granitoid rocks.

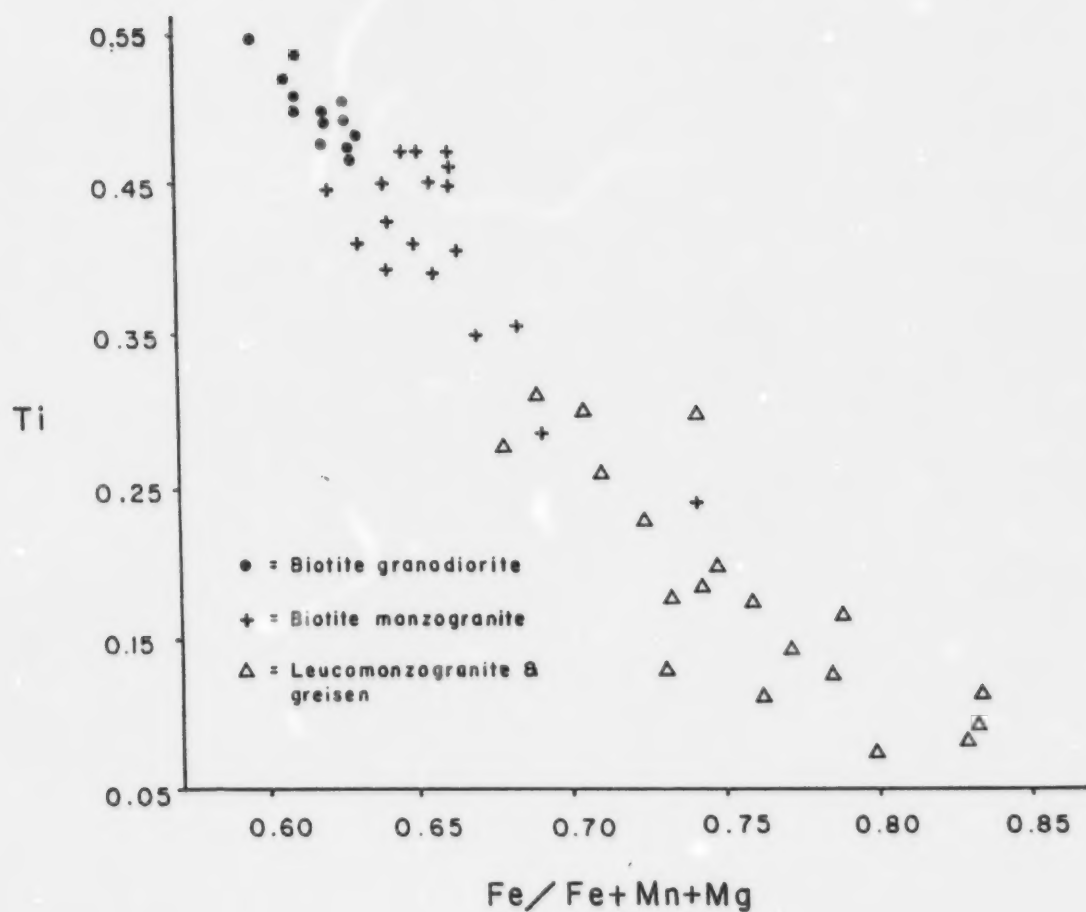


Figure 9-6. Ti vs. Fe/(Fe+Mn+Mg) plot for biotite from the Davis Lake Complex. ● granodiorite; + biotite monzogranite; + muscovite-topaz-biotite monzogranite; △ quartz-topaz greisens.

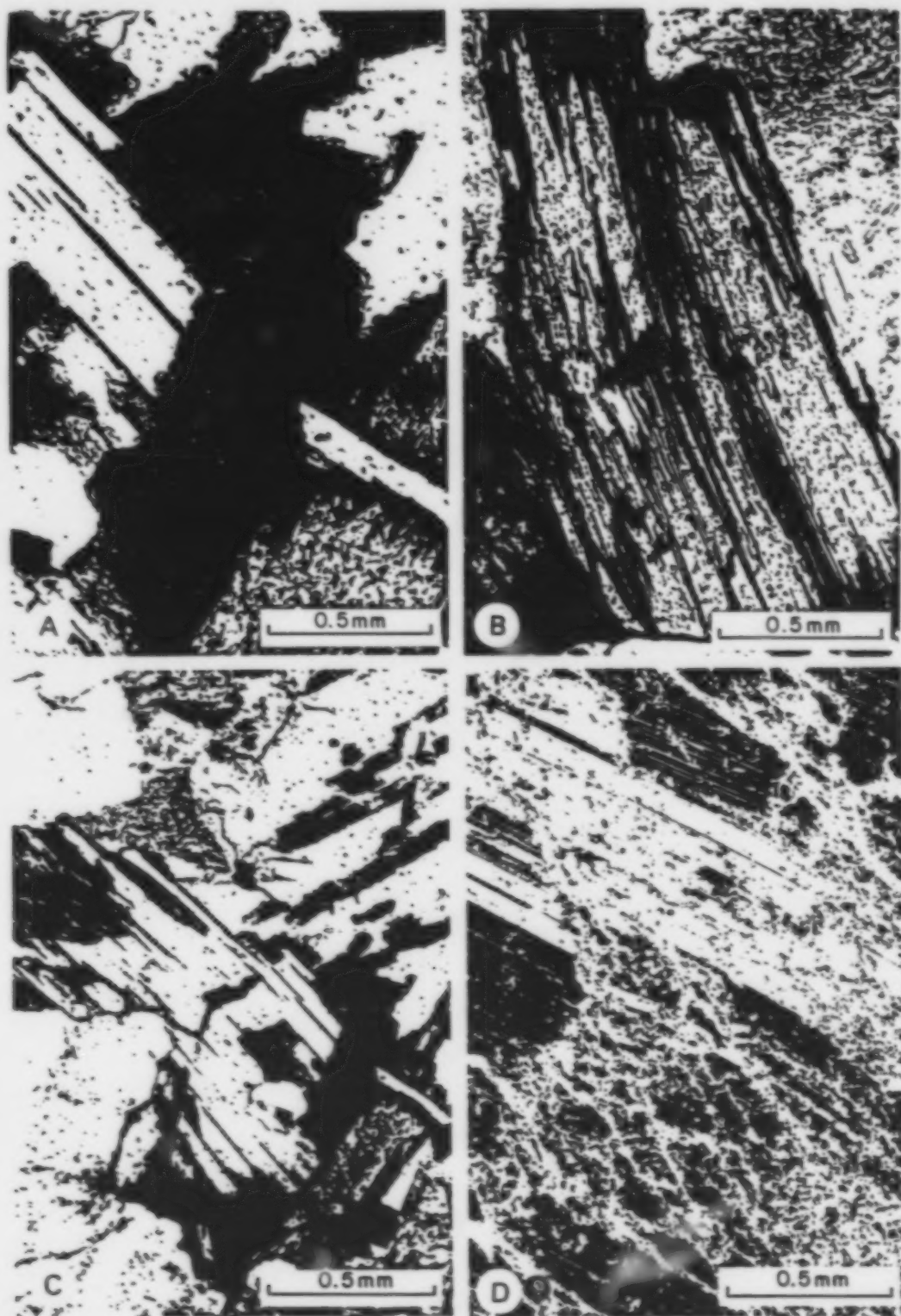


Figure 9-7. Photomicrographs illustrating various stages of muscovitization.

(Mg,Fe)Si Al^{IV}Al^{VI}. McNamara (1965) showed that deviation of muscovite composition from the muscovite-celadonite join is essentially due to replacement other than the above substitution. The composition of muscovite from the Davis Lake Complex is plotted in a A-S-F diagram (Figure 9-9) and representative analyses of muscovite from various rock types is given in Table 9-2.

FLUORITE

Fluorite is widely distributed through the granitoid rocks of the Davis Lake Complex and it generally makes up less than one per cent of the rock. It is mostly white but also pale green and pale to deep purple, and the colour varieties occur both separately and together. Typically, it occurs as vug filling in silicified granite (Figure 9-10), as discrete grains with topaz along the twin lamellae in albite, with cassiterite along cleavage planes in biotite and muscovite (Figures 9-11) and locally as cubic crystals in the quartz-topaz-cassiterite veins. Small crystals are seen, with the microscope, in or adjacent to cassiterite, chalcopyrite and sphalerite, and locally it is found intergrown with muscovite and chlorite. Some fluorite crystals are phosphorescent while others are not, perhaps indicating separate episode of deposition or slightly different trace element composition.

The importance of fluorine as a complexing and mobilizing agent for rare earth elements, uranium and tin in the SMB has been discussed by Muecke and Clarke (1981) who pointed out the potential of using REE contents of fluorites as an exploration tool. The REE contents, together with other trace elements of six fluorites, is given below:

	1	2	3	4	5	6	F1*
La	3.80	2.89	3.56	3.75	3.20	2.90	2.75
Ce	12.16	12.45	13.16	11.80	11.39	10.87	11.30
Nd	10.28	11.20	12.00	9.78	10.40	10.58	10.20
Sm	5.21	6.30	5.58	4.58	4.73	5.30	4.98
Eu	1.21	0.98	1.16	1.87	0.66	0.95	1.67
Tb	1.39	1.46	1.80	1.65	1.85	1.35	1.85
Yb	6.65	5.32	6.90	8.28	5.95	7.50	7.38
Lu	0.83	0.85	0.92	0.63	1.05	0.58	0.97
Ba	415	330	405	360	385	444	395
Ca	-	-	-	-	-	-	0.10
U	20.6	26.6	33.6	42.5	29.8	30.0	-
Th	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	0.11
Hf	0.17	0.18	0.14	0.11	0.19	0.14	0.15

* Data from Muecke and Clarke (1981)

From the above data it is clear the fluorite is characterized by high REE and, in two cases, exceeds the REE content of the whole rocks. Other characteristics are high Ba and U contents, and low Th and Hf contents.

The chondrite normalized patterns are shown in Figure 9-12a, b. These patterns are remarkably similar to that established by Muecke and

Clarke (1981) for the fluorite associated with the mineralization.

CASSITERITE

Cassiterite has been found in porphyritic biotite-granite, altered granite, albitites and albitized granites, quartz-muscovite greisen and quartz-topaz greisen. In each of these, the cassiterite has distinctive physical characteristics which probably reflect slightly different stages of deposition. The details are discussed in the following paragraphs and illustrated in Figure 9-13.

Cassiterite from the biotite granite is generally found as inclusions in biotite flakes and seldom as granular aggregates. The grains have an average 0.1 mm diameter across, are predominantly magnetic, are intensively pleochroic ranging from blood red to dark brown, and are devoid of twinning but showing color zoning. The cassiterite from the altered granite is weakly pleochroic, non-magnetic, and slightly purplish but generally light to drab brown. The crystals are devoid of zoning but are twinned on (011), as readily observed in thin section.

The cassiterite from albitized granites and albitites is typically composed of two types within the same thin section, i.e. a pale yellow to dark brown type and a colourless non-pleochroic type. The dark brown variety occurs as highly corroded grains, and may be replaced by quartz and muscovite. The colourless variety is a later generation and occurs predominantly as multiply-twinning discrete grains and along fractures in the dark brown variety. Similar textures have been described by Hoeking (1970) and Greaves et al. (1971). Often the colourless cassiterite is associated with pyrite, pyrrhotite, sphalerite and chalcopyrite.

The cassiterite from greisens occur as finely disseminated grains up to 0.5mm in diameter, and as coarse grained masses in quartz veins. In rare cases, cassiterite masses show imperfect, short, prismatic or pyramidal forms. Frequently, the greisen cassiterite is composite grains, highly variable and corroded in appearance, and shows evidence of physical transportation. Generally, the pleochroism is located erratically within specific areas of individual crystals. The zoning commonly takes the form of a central dark area with any number of alternating bands around it. Almost identical features were described by Moore and Howie (1979) for Cornish cassiterite. However, the most common twinning for Nova Scotia cassiterite is cyclic whereas for Cornish cassiterite it is polysynthetic. Cassiterite in the greisens is typically associated with chalcopyrite, sphalerite, pyrrhotite, tetrahedrite and native bismuth. In many cases, sulphides rim and replace the cassiterite.

The cassiterite from various geological environments was analyzed for trace elements, since similar studies elsewhere (Borisenko and Lizunov, 1959; Stumpf, 1963; Schneider et al. 1977, 1978) suggested that the chemistry of cassiterite had considerable potential in tracing

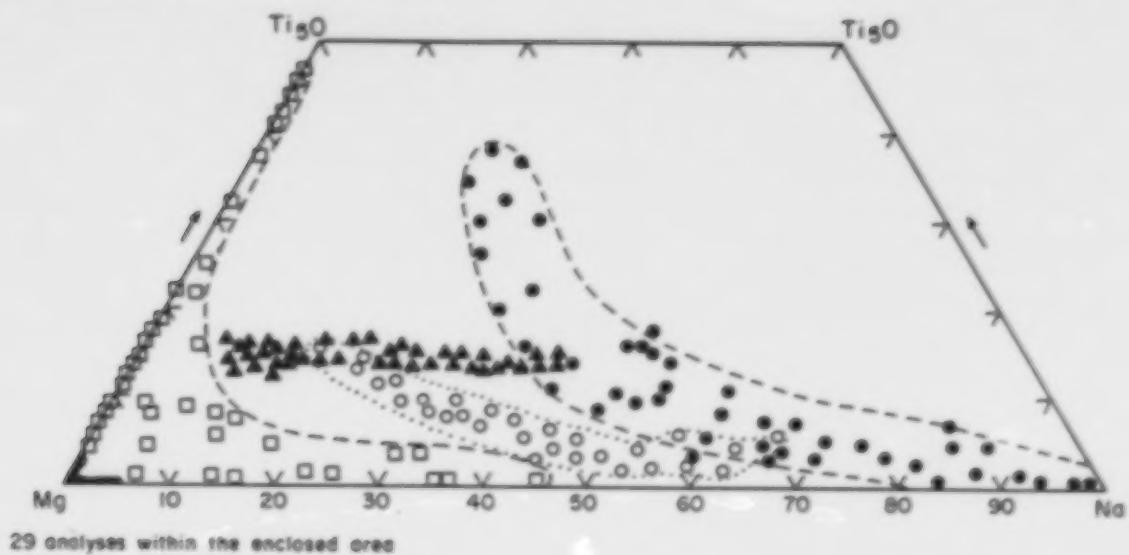


Figure 9-8. Mg-Ti-Na (cation proportions) diagram showing muscovite composition from the East Kemptville($\square$). Muscovite composition from other Sn-W-U-Mo-Cu deposits in Nova Scotia is shown for comparison. Cu-Mo porphyry deposit at Coxheath; $\circ$ W-Mo-Sn deposit at Long Lake; $\circ$ U deposit at Millet Brook.

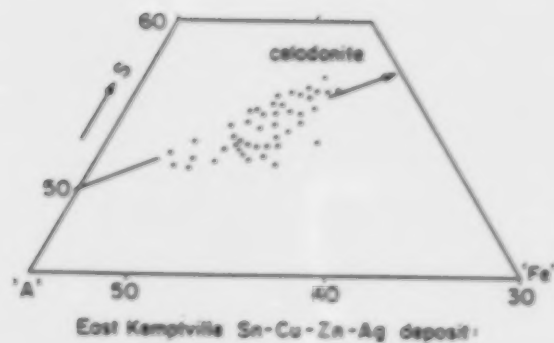


Figure 9-9. Al-Si-Fe-(Fe-Mn-Mg) diagram showing the composition of muscovite from granitoid rocks and quartz-topaz greisen.

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Table 9-2. Representative microprobe analyses of muscovite from various rock types from the Davle Lake granitoid rocks.

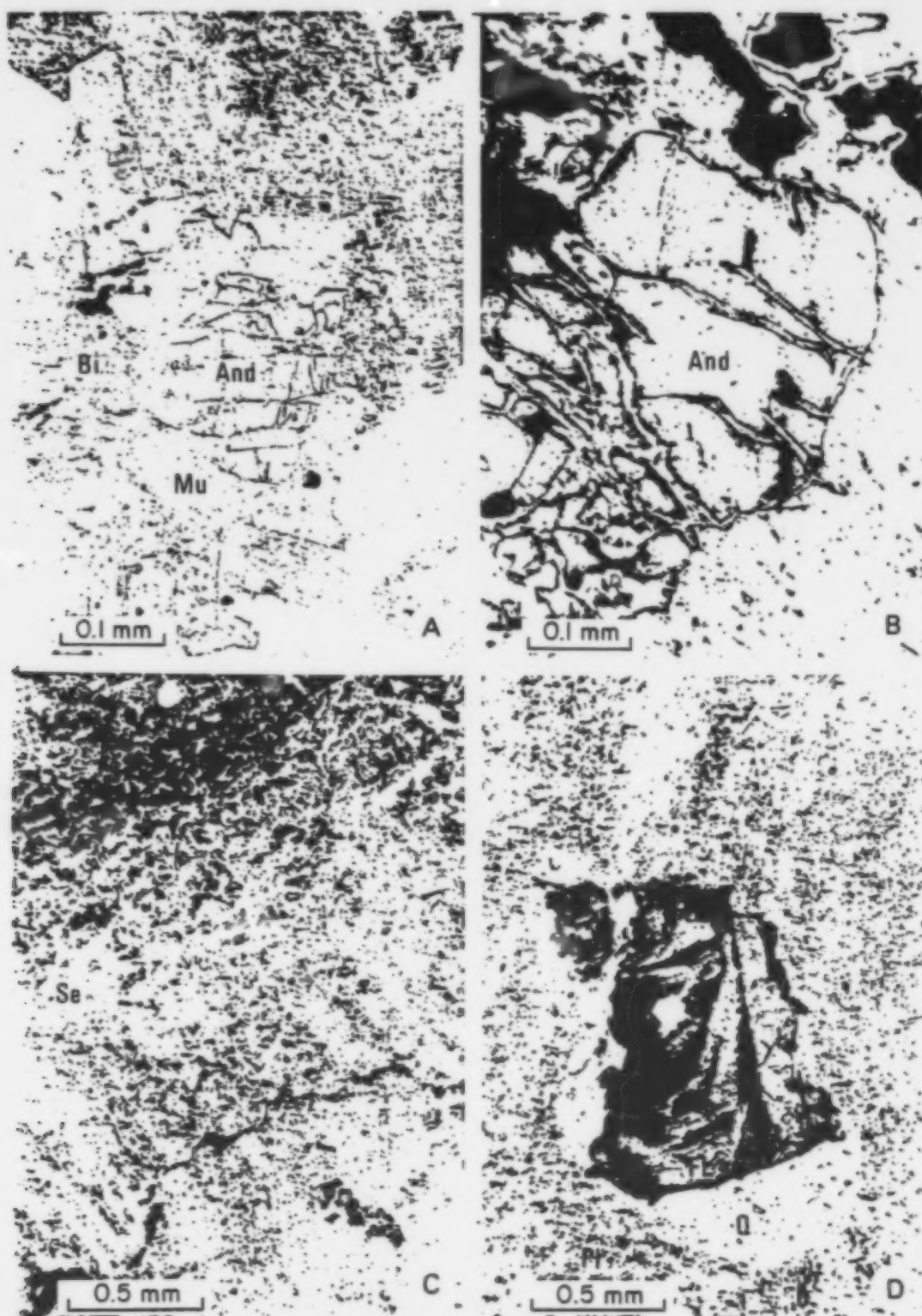


Figure 9-10. Photomicrographs illustrating the muscovitization of andalusite, topaz and fluorite during muscovitization. Note the presence of feldspar and quartz in vugs.

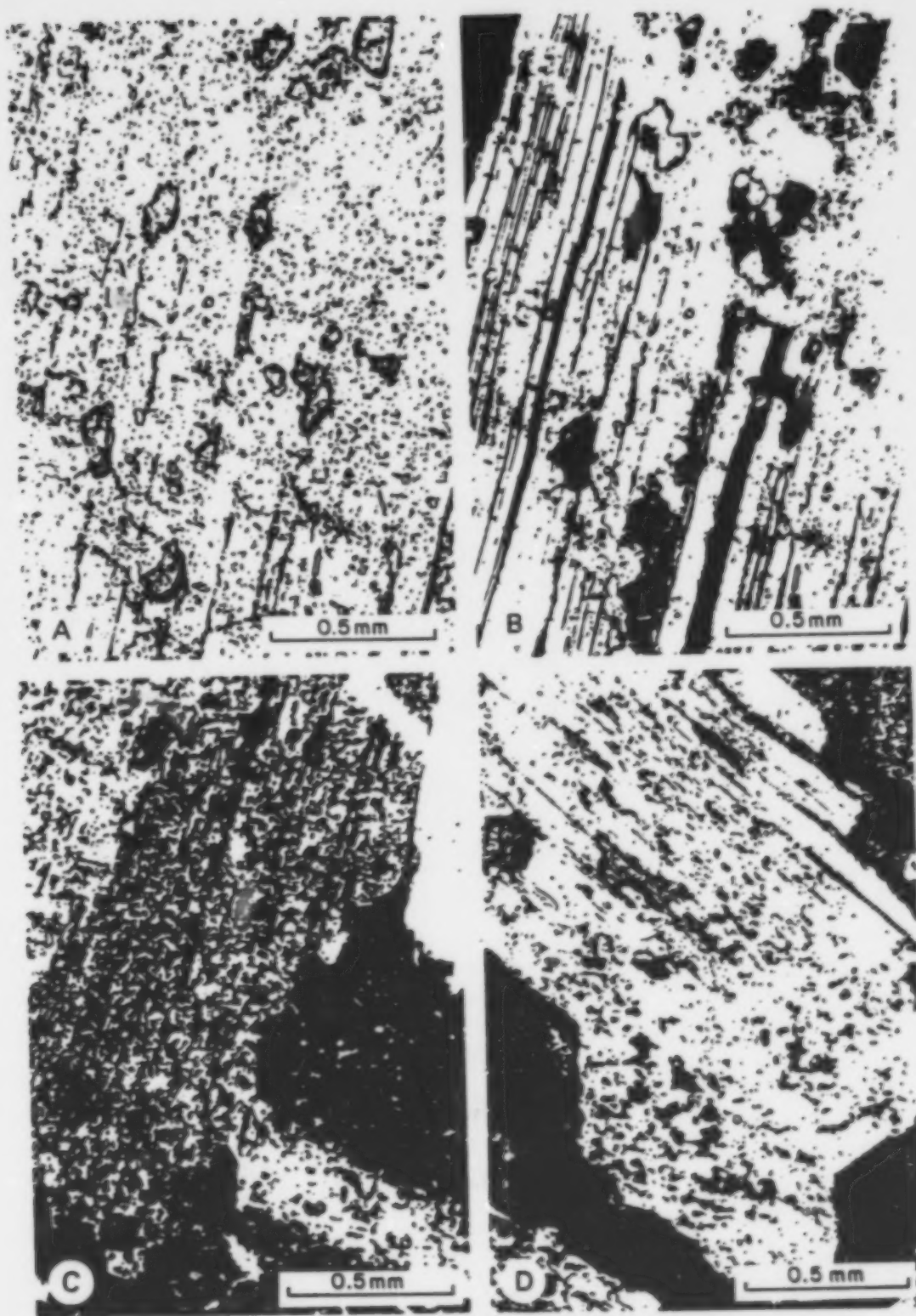


Figure 9-11. Photomicrographs illustrating the fluoritization of albite, biotite and muscovite. Note the presence of fluorite and topaz along the twin planes in albite.

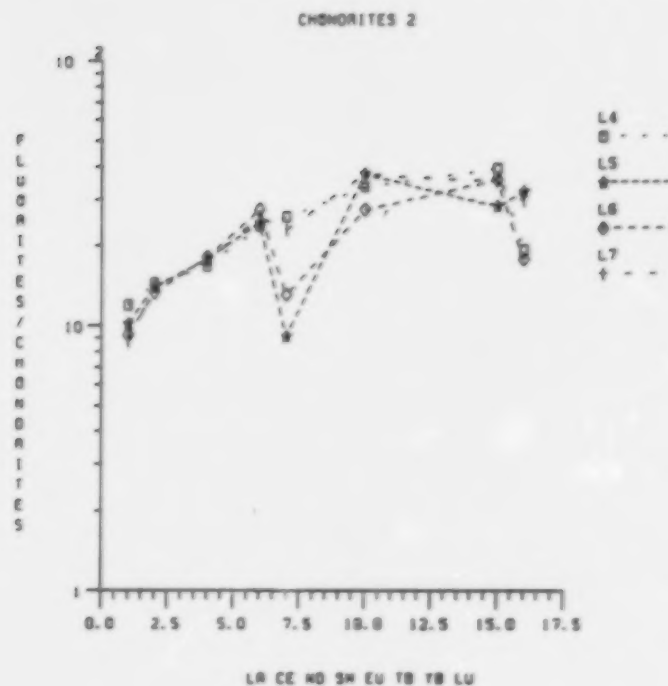
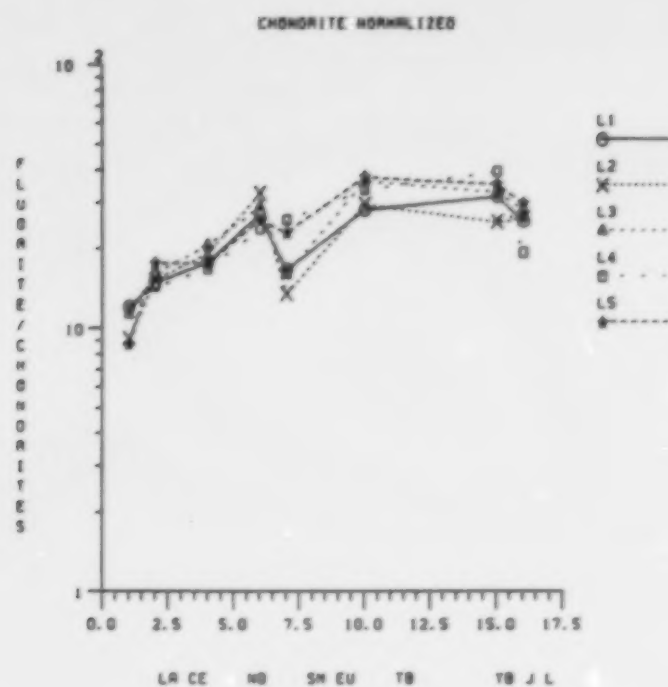


Figure 9-12. Chondrite normalized REE patterns for fluorite from the Davis Lake granitoid rocks. The greisen fluorite are numbered 1 and 2 and the remaining patterns are from the muscovite-quartz-topaz granite.

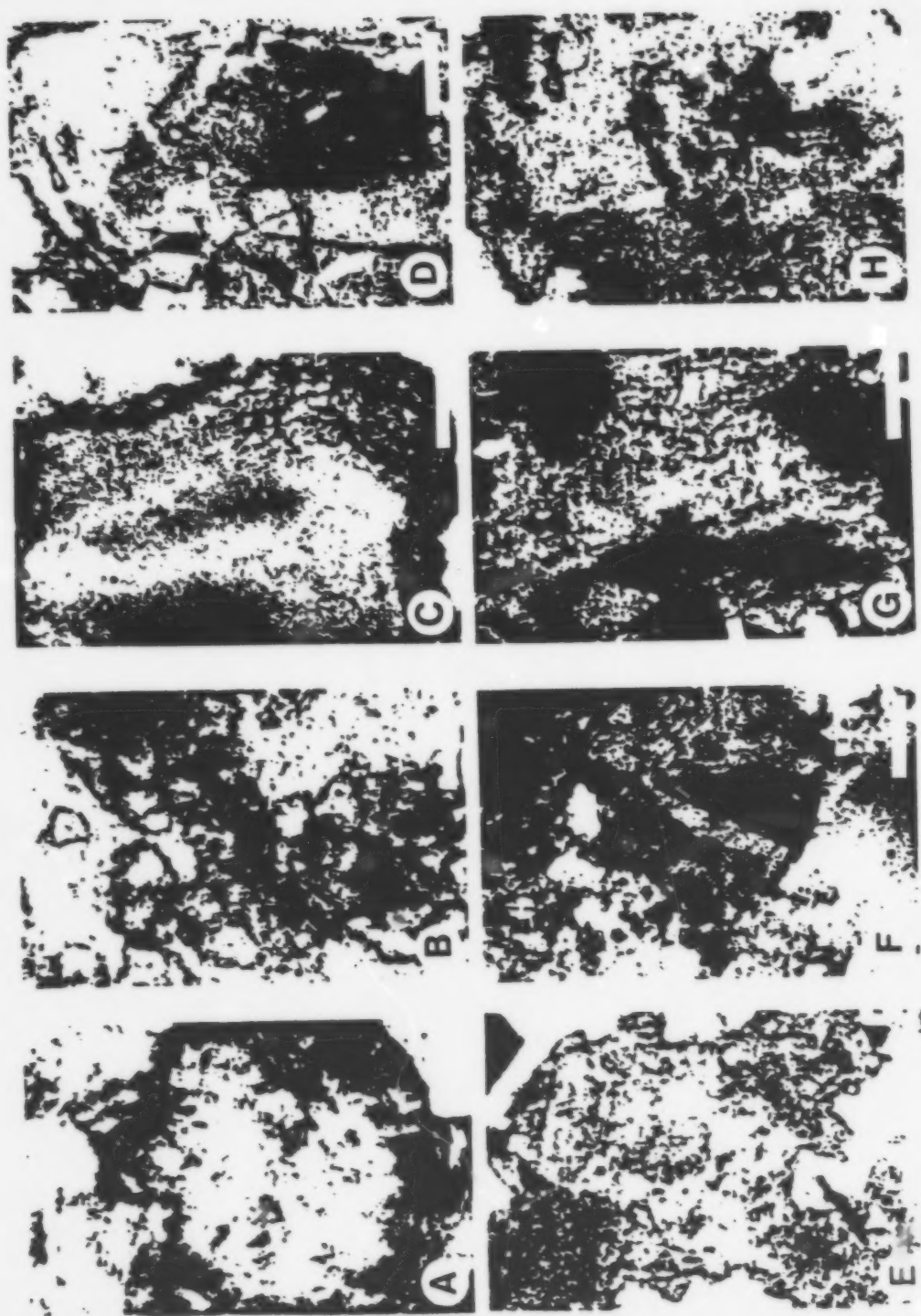


Figure 9-13. Photomicrographs showing the twinning, zoning and replacement of cassiterite by sulphides from various rock types from the Davis Lake area. (See text for explanation).

Table 9-3. Composition and cell volume of cassiterite from various environments in southwestern Nova Scotia.

No.	Sn	Fe	Al	Ti	Ta	Nb	Bi	In	Fe	Al	Ti	Ta	Nb	Bi	Atom proportion (%)			E (mV)	Impurity	Cell volume (μm^3)	Enviro- nment
															Ti-Ta-Nb	Ti	Ta	Nb			
1	77.45	10.44	0.37	0.14	0.07	2.87	0.12	0.6550	0.0158	0.0020	0.0029	0.0003	0.0308	0.0090	0.0311	9.32	0.76	89.72	7.967	71.274	1
2	78.92	0.14	0.01	0.22	0.06	3.12	0.07	0.6623	0.0025	0.0000	0.0005	0.0003	0.0335	0.00030	0.0383	11.74	0.78	87.48	6.2076	71.345	1
3	76.92	1.67	0.01	0.85	0.10	3.50	-	0.6463	0.0298	0.0000	0.0177	0.0005	0.0378	-	0.0558	31.72	0.89	67.39	13.2448	71.298	1
4	77.77	0.56	-	0.14	0.10	0.63	0.01	0.6535	0.0100	-	0.0029	0.0005	0.0067	0.00004	0.0101	28.71	1.45	66.34	16.8600	nd	1
5	77.51	0.14	0.44	0.14	0.05	1.37	0.63	0.6496	0.0025	0.0020	0.0039	0.0002	0.0147	0.00300	0.0188	20.74	1.06	78.20	4.0948	71.335	1
6	78.80	0.11	-	0.13	0.09	0.26	0.02	0.6621	0.0019	-	0.0027	0.0004	0.0038	0.00009	0.0069	29.13	0.79	55.08	1.3291	nd	1
7	78.67	0.73	0.32	0.04	0.05	0.19	0.01	0.6610	0.0130	0.0010	0.0018	0.0002	0.0020	0.00004	0.0060	47.00	5.00	50.00	1.3161	nd	1
8	77.48	0.10	0.64	0.06	0.03	0.17	0.24	0.6552	0.0017	0.0010	0.0012	0.0001	0.0018	0.00110	0.0031	38.70	1.22	58.08	1.2134	nd	1
9	77.30	1.20	0.06	0.95	1.05	1.85	-	0.6495	0.0214	0.0000	0.0197	0.0001	0.0148	-	0.0269	40.28	18.60	41.12	10.8237	71.278	1
10	77.44	0.32	1.05	0.11	0.29	0.65	0.41	0.6568	0.0080	0.0000	0.0022	0.0021	0.0069	0.00180	0.0112	19.64	18.75	61.61	3.8470	71.260	1
11	77.85	0.04	-	0.11	0.54	1.16	0.15	0.6626	0.0007	-	0.0022	0.0030	0.0124	0.00070	0.0176	12.50	17.00	70.50	2.6667	71.290	1
12	78.85	1.08	0.07	0.11	1.35	3.20	-	0.6457	0.0192	0.0000	0.0022	0.0075	0.0344	-	0.0441	4.95	17.50	78.05	0.2740	nd	1
13	77.25	0.16	-	0.08	0.93	0.73	0.04	0.6500	0.0028	-	0.0016	0.0051	0.0078	0.00019	0.0145	11.03	35.17	53.80	2.6700	nd	4
14	77.63	0.53	0.83	0.12	0.43	0.54	1.66	0.6523	0.0094	0.00050	0.0025	0.0036	0.0058	0.00794	0.0119	21.00	30.25	48.75	5.1600	71.352	4
15	79.10	0.55	1.61	0.05	0.05	0.27	0.06	0.6638	0.0096	0.00070	0.0010	0.0002	0.0029	0.00020	0.0061	24.39	48.78	26.83	3.4347	71.365	4
16	77.42	1.40	0.11	0.12	0.84	0.04	0.04	0.6547	0.0250	0.00060	0.0025	0.0044	0.0004	0.00019	0.0073	34.24	60.27	5.49	0.0500	nd	5
17	77.01	0.02	-	0.16	1.65	-	0.03	0.6555	0.0003	-	0.0033	0.0091	-	0.00010	0.0124	26.61	73.38	0.01	1.9527	71.478	5
18	78.23	0.61	-	0.27	4.39	-	-	0.6573	0.0108	-	0.0056	0.0243	-	-	0.0299	18.72	81.27	0.01	6.1919	71.320	5
19	78.40	0.12	-	0.29	11.80	0.02	0.07	0.6588	0.0021	-	0.0080	0.0655	0.0002	0.00030	0.0717	8.36	91.35	0.29	11.2400	71.310	5
20	78.89	0.97	-	0.09	0.17	0.02	-	0.6629	0.0173	-	0.0018	0.0009	0.0002	-	0.0029	62.06	31.03	6.91	3.0170	71.372	6
21	78.35	0.11	-	0.16	0.43	0.03	0.02	0.6584	0.0019	-	0.0033	0.0023	0.0003	0.00009	0.0059	55.93	28.98	5.09	1.1983	71.505	6
22	77.58	0.39	-	0.44	0.54	0.02	0.01	0.6519	0.0069	-	0.0091	0.0030	0.0002	0.00004	0.0123	73.98	24.39	1.63	2.9500	nd	6
23	78.31	0.36	-	0.48	-	0.01	-	0.6580	0.0069	-	0.0090	-	0.0001	-	0.0101	99.00	0.00	1.00	2.5800	nd	7
24	77.65	0.29	-	0.35	0.08	0.01	0.01	0.6525	0.0052	-	0.0072	0.0004	0.0001	0.00004	0.0077	93.50	5.19	1.31	2.0300	71.450	7
25	77.79	0.60	-	0.34	0.14	0.02	-	0.6536	0.0107	-	0.0070	0.0007	0.0002	-	0.0079	88.60	8.86	2.54	2.8600	nd	8
26	77.35	0.64	-	0.77	0.67	0.02	0.01	0.6500	0.0114	-	0.0160	0.0037	0.0002	0.00004	0.0199	80.40	18.59	1.01	4.8200	nd	8
27	78.45	0.51	0.17	0.74	0.80	0.04	0.04	0.6592	0.0091	0.00090	0.0154	0.0044	0.0004	0.00019	0.0202	76.23	21.70	2.07	4.6100	71.355	8
28	78.67	0.52	-	0.51	0.73	0.16	0.08	0.6610	0.0092	-	0.0106	0.0040	0.0017	0.00038	0.0163	65.03	24.53	10.44	3.9100	nd	8
29	78.42	0.30	-	0.15	0.12	0.02	0.01	0.6589	0.0053	-	0.0031	0.0006	0.0002	0.00004	0.0039	79.48	5.38	5.14	1.3952	nd	8
30	78.00	0.64	-	0.73	0.38	0.02	0.01	0.6554	0.0114	-	0.0152	0.0021	0.0002	0.00004	0.0175	86.95	12.00	1.15	4.6095	71.372	8
31	78.98	0.21	-	0.26	0.03	0.04	-	0.6636	0.0037	-	0.0054	0.0001	0.0004	-	0.0059	91.52	1.69	6.79	1.4466	nd	9
32	78.93	0.49	-	0.29	0.01	0.03	0.01	0.6632	0.0087	-	0.0060	0.00005	0.0003	0.00004	0.0063	95.23	0.79	3.98	2.2753	71.425	9
33	78.01	0.60	-	0.70	0.05	0.26	0.21	0.6555	0.0107	-	0.0145	0.0002	0.0027	0.00100	0.0174	83.33	1.14	15.53	4.4393	nd	9
34	78.53	1.26	1.21	1.06	0.09	0.32	0.27	0.6599	0.0242	0.00650	0.0220	0.0005	0.0034	0.00120	0.0259	84.94	1.93	13.13	1.6820	71.479	9
35	78.01	0.77	0.68	1.19	0.83	0.67	2.20	0.6555	0.0137	0.00720	0.0247	0.0048	0.0072	0.01050	0.0365	67.67	12.60	19.73	10.3500	71.299	9
36	78.41	0.18	-	0.18	0.04	0.09	0.01	0.6589	0.0032	-	0.0037	0.0002	0.00096	0.00004	0.0049	76.13	4.11	19.76	1.2200	71.498	9
37	78.30	0.21	-	0.28	-	0.03	-	0.6579	0.0037	-	0.0058	-	0.0003	-	0.0061	95.08	0.00	4.92	1.4800	nd	9
38	78.12	0.75	-	0.47	0.22	0.04	-	0.6564	0.0130	-	0.0097	0.0012	0.0004	-	0.0113	85.84	1.06	13.10	3.7000	nd	9
39	78.40	0.48	0.21	0.11	0.05	0.04	0.01	0.6630	0.0014	0.00005	0.0022	0.00027	0.0004	0.00004	0.0028	78.57	9.64	11.79	0.6700	71.548	9

the genetic history of tin mineralization. The results are given in Table 9-3. It was found that cassiterite which is rich in Fe, W and Bi has discrete inclusions of magnetite, wolframite and native bismuth. This means that no genetic significance could be attached to the cassiterite with respect to these elements because it could be seen that the solution which deposited cassiterite resulted in the incorporation of pre-existing phases.

Discrete mineral phases of Nb, Ta and Ti were not observed in the cassiterite of southwestern Nova Scotia, hence the concentration of these elements is taken to indicate cassiterite compositions, which are plotted in a Nb-Ta-Ti ternary diagram in Figure 9-14. It is evident that the cassiterite of the porphyritic biotite granite is Nb-rich, whereas the cassiterite of altered granites is slightly enriched in Ta and depleted in Nb with respect to biotite granite. The cassiterite from albitized granites, albitites and greisens are all deficient in Nb, although the cassiterite from albitized granites is Ta-rich and Ti-poor, and the greisen cassiterite is Ti-rich. On the basis of this variation and within the limits of chemical and sample uncertainties, a generalized trend for the evolution of various cassiterites in the granitoid rocks of the Davis Lake Complex is shown in Figure 9-14.

For comparison, the Nb:Ta:Ti proportions in cassiterite from the pelitic rocks of southwestern Nova Scotia are also shown in Figure 9-14. It is evident that the evolution of cassiterite in the metasediments follows a remarkably similar trend. Because of this similarity in trend it is tempting to propose that the deposition and evolution of tin mineralization in the metasediments and the granitoid rocks resulted from similar fluids and genetic controls.

Goncharov and Filatov (1971) suggested that the conditions of formation of cassiterite could be correlated with cell parameters. The cell volumes were determined for only a few cassiterite from southwest Nova Scotia (Table 9-3) and when these are plotted against total impurities, it was found that the cell volumes markedly decrease with increasing impurity content, a finding which is at variance with Wallis (1971) and in accord with Clark et al. (1976). This plot also demonstrates that the greisen cassiterite is generally poor in total impurities, whereas the reverse is true for all the remaining rock types.

WOLFRAMITE

Small concentrations (<1%) of wolframite occur in cores of quartz-topaz veins and rarely large crystals (>3 cm) have a stubby tabular to prismatic habit, are generally subhedral to euhedral, and occur either singly or as composite clusters. The wolframite is dark brownish black on fresh cleavage surfaces, showing deep-blood-red internal reflection. Cleavage is well developed on (010), and cleavage (parting) less developed on (100). Thin quartz-topaz veins (observed in DDH79-35) rarely display small

necoles of prismatic crystals that grow perpendicular into the vein from its northwest wall. In this morphological form crystals commonly fan or flare outward from their base in the direction of growth. The centre of the base of such a crystal is in direct contact with the wall and, outwardly, the crystal rests on a progressively thicker selvage of muscovite, which shows thickest development on the walls between wolframite crystals. This relationship indicates simultaneous deposition of wolframite and muscovite beginning with the onset of vein mineralization. Inclusion of flakes of muscovite within wolframite also indicates synchronous crystallization of wolframite and muscovite.

Many of the subhedral to euhedral crystals of wolframite show growth zones that, as seen in thin sections, are alternating lighter and darker brown zones that reflect the crystallographic form of the growing crystals. Some crystals have been fractured, dismembered, or otherwise deformed during growth and these growth zones are often filled with chalcopryrite, sphalerite, quartz and topaz. The dismemberment resulted from brecciation rather than replacement, as shown locally by segments of wolframite that can be fitted together after removal of interstitial materials.

Microprobe data were acquired to establish the iron-manganese compositions, and compositional variations, if any, in visible growth zones, for the purpose of better defining the environment of their composition. The analytical data show iron content of the wolframite ranges from about 0.6 to 14.35 weight per cent. In general it was found that the highest Mn/Fe ratio was characteristic of wolframite associated with other sulphides and the ratio was low when wolframite was not associated with sulphides.

The significance of variations in manganese and iron content of wolframite, commonly stated as the huebnerite-ferberite molecular ratio (H/F), as an indication of condition of deposition, is controversial. The detailed discussion by Delaner (1952, 1954), Bolduan (1954), Varlamoff (1958) and de Magnee and Aderca (1960) indicated that huebnerite has been deposited at higher temperatures than ferberite. However, studies by Ganeed and Sechina (1960), Taylor and Hosking (1970) presented data to suggest that the opposite has been true. Data of Clark (1970) on the Cornwall deposits indicated that wolframite in pegmatitic veins contained more iron than that in associated "normal lodes". Moore and Howie (1978) indicated that uneven distribution of manganese and iron may be controlled not only by the concentration of these elements in ore fluids but by rapid changes in the composition of ore fluids, by competing mineral phases, and by subsequent remobilization of manganese and iron. Oscillatory zoning of iron and manganese in wolframite crystals may be controlled by local concentration gradients at the crystal-liquid interface, or by periodic supercooling of the liquid, or by "constitutional" supercooling as proposed for oscillatory zoning in plagioclase feldspar by Sibley et al. (1976), or the thermodynamic bychanges proposed by Amosse (1980). The wolframite from East

Kemptville deposit show marked changes in the H/F ratio and this is observed in a single thin section or portion of a grain where the grain is in contact with contrasting coexisting phases. Though our data is of preliminary nature, we conclude that H/F is of little value as a clue to temperature, pressure and oxygen and sulphur fugacity during the mineralization episode.

PYRITE

Pyrite occurs in all granitic rock types of the Davis Lake Complex as minute inclusions in biotite and intergranular aggregates. Minute aggregates also occur in, on or next to silicate minerals which show evidence of sericitic and chloritic alteration. These aggregates are found in fractures, along cleavage and twin planes, and along silicate grain boundaries. Pyrite often has inclusions of chalcopyrite and pyrrhotite.

Electron analyses for Fe, S, Cu, Ni and Co were carried out on 11 polished thin sections from (1) unaltered biotite monzogranite (2) hydrothermally altered rock associated with the quartz-topaz greisen and (3) quartz-topaz greisen. The results show the Fe content is highest in biotite monzogranite and lowest in the greisen samples. Sulphur values are lowest in the monzogranite and similar in hydrothermally altered rocks and greisen.

The Co:Ni ratio of pyrite has been the subject of intense study but has been mostly confined to volcanogenic massive sulphides (e.g. recent review by Bralier et al.) (1979). Twenty-seven samples were analysed for Co and Ni, and it is found that Co and Ni contents as well as the Co/Ni ratio of the host rocks are strongly correlated. This correlation is true for most granite rocks. However, in the tin-rich greisens, pyrite has a highly variable content of Co and Ni as well as a highly variable Co:Ni ratio. This variability is present in pyrites in the same hand specimen. However, the mean Co:Ni ratio of pyrite from East Kemptville is almost identical to the values reported for Butte ore.

Many samples of hydrothermally altered rocks contain chlorite and pyrite, and this pair shows a positive correlation, whereas in the unaltered rocks pyrite is associated with biotite and the pair shows a negative correlation. The data, though limited, would suggest that during hydrothermal alteration involving copper-rich solution, newly formed chlorite and pyrite takes copper and presumably at magmatic temperatures with little copper available, most copper would enter pyrite. Our result also suggests that pyrite forms more or less continuously from the magmatic through to the greisenization stage.

OTHER ASSOCIATED MINERALS

Two varieties of tetrahedrite have been identified in polished thin sections from chalcopyrite-rich greisen. Tetrahedrite 1 forms anhedral grains that have a more or less constant size of about 0.1mm and are characterized by numerous fine cracks. Microprobe analyses have shown that this tetrahedrite is low in Ag (Ag, 1.4%; Cu, 42.6%; Sb, 31.7%; and S, 23.9%).

Tetrahedrite 2 (Ag, 10.4%; Cu, 38.6%; Sb, 25.3%; and S, 24.7%) forms irregular patches and generally replaces tetrahedrite 1. Polybasite is rare, occurring as anhedral grains associated with tetrahedrite. Chalcostibite is found in close association with polybasite and has a composition of 1.01% Cu, 0.04% Ag, 1.00% S and 1.95% Sb, which is almost identical with that of ideal chalcostibite. The beryllium minerals phenakite and bertrandite are identified in the albitized granite and albitite breccia. The phenakite occurs as disseminated subhedral crystals and the bertrandite as fine-grained intergrowths with albite and muscovite.

GEOCHEMISTRY OF THE GRANITOID ROCKS AND GREISENS

MAJOR AND TRACE ELEMENT GEOCHEMISTRY

The analytical data together with the CIPW norms for the granitoid rocks of the East Kemptville area and the quartz-topaz greisen are given in Tables 9-4 and 9-5, respectively, and the statistical summary is presented in Tables 9-6 and 9-7.

As shown in Figure 9-15, the porphyritic granites, like those of most of the Sn-bearing granites, plot within the field of typical granites. The effects of albitization are clearly progressive from the granites toward albitites, with some degree of K-metasomatism in highly altered granites. In general, however, the alteration trend reflects variation of either albite or quartz components, with relatively little variation in orthoclase. The greisenization trend leads from the fresh granites through sericitization to greisen and quartz greisens. Although the covariance of other elements could be evaluated in terms of variation diagrams (c.f. Richardson, 1985), interpretation difficult and unless care is taken, may lead to conflicting trends, even though these trends may be due to cumulative effects of various magmatic and postmagmatic processes. Due to these reasons we have chosen to summarize the major and trace element data in terms of the matrix of correlation coefficients. As shown in the statistical summaries, the geochemical data depart from the typical normal distribution. Hence the entire data set was log transformed to compensate for the skewness. The R-matrix for the granitoid rocks and the greisen is presented in Figures 9-16 and 9-17 respectively. From these granite Sn in the granite is positively correlated (statistically significant) with Na₂O, and negatively with Fe₂O₃, FeO, MnO, MgO and in the greisen it is positively correlated with P₂O₅, B, W and Al₂O₃ and negatively with F, Sr and U. Another significant feature is in the granitoids, uranium is not correlated with any of the major and trace elements.

The contrasting behaviours of some of the characteristic trace elements in the granitoid rocks and the greisens are illustrated in Figure 9-18 to 9-22. Two elements most characteristic of greisens, Li and F, are plotted against Sn for both the granitoid rocks and greisens, showing that a relatively constant Li/F ratio is maintained during Sn-enrichment though the mean

values of Sn in the granite and greisens are vastly different (see Tables 9-6 and 9-7). Figure 9-18 also shows that the Li/F ratio reaches slightly lower values in some granite samples and there is a complete overlap between the granite and greisen samples throughout the full Sn-enrichment trend. This might be explained by suggesting Sn-mineralization was essentially a magmatic process, with a continuous transition from granite to greisen. It certainly indicates a similarity in the geochemical behaviour of Li and F in both granitic melts and greisen fluids. Figures 9-19 to 9-20 show there is an increasing separation of the granite and the greisen samples from Li/F through Li/U to Li/Th ratios throughout the full range of Sn enrichment, reflecting the differing geochemical behaviour of F, U and Th in the granitic melts and greisen fluids. Great contrasts in the geochemical behavior of different elements in the granitic melts and greisen fluids can be inferred from the different ratios seen in Figure 9-19 and 9-20. Of particular note are the higher proportions of Th and Zr relative to most elements in the granites compared to the greisens, resulting from the magmatic control on Th and Zr, and their immobility in the magmatic hydrothermal greisen fluids.

RARE EARTH GEOCHEMISTRY

The REE geochemistry from the East Kemptville tin deposit was discussed in detail by Chatterjee and Strong (1984a) who showed with increasing stages of alteration, as reflected by decreasing SiO_2 concentrations, there does seem to be a consistent change from high-Si to low-Si samples, namely a decrease in LREE concentration, a decrease in size of the europium anomaly, and a slight increase in HREE. These are all features which would be predicted for alteration and for deposits resulting from fluids containing Cl (LREE-leaching), F and CO_3 (HREE-deposition).

The REE patterns for greisens when arranged in order of increasing Sn concentration show the lowest Sn is 180 ppm in the albitized leucomonzogranite which shows low total REE concentration and a relative depletion in HREE. The albitite sample although having a similar Sn concentration of 187 ppm, displays the reverse REE pattern, with depletion of LREE and enrichment in HREE. It may be significant that this coincides with a drop in SiO_2 to 64.69% and stronger albitization (58.60% normative Ab). With further albitization (e.g. sample 386 with 60.96% normative Ab) and slight increase in Sn (230 ppm), the LREE are virtually totally depleted and the HREE are also somewhat decreased. With greisenization and further Sn-enrichment, the LREE remain depleted and the HREE are more or less completely removed.

Chatterjee and Strong (1984a) interpreted these REE variations in terms of fluid chemistry, i.e. transport and deposition by Cl, F and CO_3 complexes, since carbonates, fluorite and topaz are common in the East Kemptville area. Muecke and Clarke (1981) also suggested that during the cooling history of the Batholith, fluid transport of REE as fluoride complexes became increasingly important relative to fractional crystallization.

This is likewise in line with the suggestion, by Alderton et al. (1980) regarding Cornish greisens, that REE mobility is enhanced by the presence of fluorine in the hydrothermal fluids although the intensity of REE depletion observed by them is substantially less than that of the East Kemptville greisens.

The chlorite normalized REE patterns are shown in Figure 9-23. The unaltered monzogranite of the Davis Lake area is characterized by a relatively narrow range of REE, a negative europium anomaly and La/Yb ratio of approximately three. These features are, in general, similar to the monzogranite to the SMB. However, the hydrothermally altered rocks have lower REE, more pronounced negative europium anomalies and flat REE patterns. The greisen veins associated with the hydrothermally altered rocks are characterized by a systematic decrease in light and intermediate REE and an increase in heavy REE. Thus the REE patterns are rotated in an anti-clockwise manner, and with intense greisenization and associated mineralization there is a complete loss of LREE and, in most tin-rich topaz greisen, a complete loss of HREE.

DISCRIMINANT ANALYSIS OF THE GRANITOID ROCKS AND QUARTZ-TOPAZ GREISEN

From the preceding section it is clear that the granitoid rocks and the associated greisens at East Kemptville show a systematic geochemical evolution, and it appears that the behaviour of many elements was similar during both the magmatic and greisenization stages. However, it appears that the behaviour of some elements such as U, Th, W, F, Na_2O and Zr was somewhat different. To test this distinction objectively, we adopted discriminant function analysis using the SPSS program. Mathematical details for discriminant analysis are given by many authors (Morrison, 1967; Overall and Klett, 1972; Lachenbruch, 1975) and are not repeated here.

Our approach was to define a single discriminant function using stepwise analysis whereby the two populations (i.e. granitoid rocks and greisens) are separated using the Wilks method, which considers the overall multivariate F-ratio for the test of differences among the group centroids. In other words, the variable which maximizes the F-ratio also minimizes the Wilks lambda, a measure of group discrimination.

In the data set (137 samples), before any variable was removed, the Wilks lambda ranged from 0.2311 to 0.9999 and the corresponding F-values were 449.1054 to 0.00258. This indicates there exists a considerable discriminating power in the variables used. To successfully discriminate between the two populations, twenty-two steps involving 20 variables were required (out of 28). The corresponding Wilks lambda and the equivalent F for each step is shown in Table 9-8, together with the list of variables included, their tolerance, F removed and the Wilks lambda after the final step No. 22. The list of variables which are excluded by the analysis is also given in Table 9-8. This exclusion is due to the fact that the F-levels were insignificant and further computation would achieve no additional

Table 9-4. Major, trace elements and CIPW normative data for the granitoid rocks from Davis Lake area.

Category	2011	2010	2009	2008	2007	2006	2005	2004	2003	2002	2001
GRAND TOTAL	75.09	76.01	74.61	76.07	79.00	70.98	77.35	76.40	77.07	79.45	74.11
Agriculture	6.14	6.17	6.11	6.11	6.10	6.15	6.12	6.10	6.12	6.04	6.11
Manufacturing	1.55	11.46	12.11	10.94	11.20	15.29	12.47	12.39	12.05	12.95	13.71
Construction	6.97	6.45	6.70	6.19	6.75	6.39	6.48	6.41	6.50	6.12	6.13
Retail	6.97	6.72	6.77	6.10	6.78	6.74	6.78	6.73	6.55	6.67	6.17
Health	0.03	0.03	0.04	0.03	0.03	0.02	0.02	0.02	0.02	0.02	0.02
Education	0.14	0.15	0.18	0.17	0.08	0.03	0.08	0.05	0.11	0.03	0.11
Social	1.25	1.22	0.67	0.00	1.20	0.00	1.11	1.20	1.25	1.01	0.74
Transportation	3.77	3.48	3.92	3.51	2.93	4.39	4.50	4.05	4.52	3.91	3.76
Other	3.25	3.77	4.21	4.21	3.74	5.54	1.86	2.18	4.01	4.01	3.41
Total	0.49	0.66	1.17	1.13	0.97	0.97	1.26	0.88	1.15	0.49	0.41
Total	0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.09	0.03	0.10
Total	0.14	0.14	0.18	0.15	0.14	0.09	0.14	0.12	0.14	0.15	0.11
Total	0.09	0.21	0.17	0.16	0.09	0.05	0.09	0.21	0.12	0.16	0.13
Total	98.75	98.55	98.77	100.75	100.26	99.49	100.26	98.74	100.03	99.24	100.12
Agriculture	39.91	40.20	34.03	37.07	46.97	22.53	42.29	42.77	41.76	37.99	42.15
Manufacturing	1.13	0.36	0.60	1.56	1.24	0.88	1.60	1.06	0.88	0.68	0.61
Construction	23.01	22.43	23.53	25.02	19.22	33.00	11.06	13.79	11.76	13.99	13.71
Retail	29.11	29.14	34.00	30.60	24.94	37.61	38.48	35.03	38.73	33.45	34.21
Health	5.66	4.91	1.24	2.46	4.50	3.11	4.08	3.99	4.58	3.29	1.86
Education	0.33	0.36	0.43	0.40	0.20	0.00	0.13	0.13	0.25	0.05	0.40
Social	1.11	1.11	2.09	1.77	1.05	0.83	0.07	0.90	0.16	1.13	1.31
Transportation	0.25	0.45	0.50	0.25	0.37	0.56	0.70	0.59	1.41	0.16	0.41
Other	0.25	0.23	0.23	0.19	0.17	0.29	0.23	0.19	0.33	0.00	0.33
Total	0.19	0.31	0.40	0.33	0.30	0.21	0.31	0.26	0.35	0.35	0.33
Total	0.49	0.49	0.28	0.34	0.23	0.12	0.21	0.49	0.28	0.35	0.25
Agriculture	58	39	81	117	63	413	45	20	34	70	34
Manufacturing	19	22	13	18	13	9	10	9	18	65	26
Construction	149	110	110	110	110	15	34	10	10	10	10
Retail	6	10	11	4	1	31	11	3	7	1	1
Health	607.0	6000	3560	4525	4230	3835	3500	4230	3695	3560	3560
Education	185	165	212	121	109	44	35	34	37	78	19
Social	7	2	3	2	2	2	2	2	2	1	2
Transportation	7	8	14	7	6	11	8	2	9	1	3
Other	7	9	13	12	4	4	4	4	4	1	4
Total	755	752	707	754	690	779	508	590	512	665	591
Agriculture	31	24	35	31	32	136	30	25	112	89	37
Manufacturing	33	34	35	55	44	90	45	47	40	31	26
Construction	27	16	18	18	15	8	4	5	43	11	16
Retail	110	44	19	10	27	19	25	23	23	29	29
Health	18	20	22	23	24	13	24	22	22	13	13
Education	5	5	5	5	5	5	5	5	5	5	5
Social	15	12	20	16	16	21	10	8	8	14	15
Transportation	35	29	40	37	41	36	36	41	43	23	29
Other	66	64	66	60	64	47	66	62	66	49	66
Total	149	77	149	149	111	62	98	96	96	16	149

[illegible]

Figure 9-5. Major trace elements and CIPW normative data for the polymetallic quartz-topaz greisen at East Keepville (f values exceeding 1000 ppm are printed as 999, will be corrected in the final copy).

SAMPLE NO	79-35-82	79-35-83	79-35-84	79-35-85	79-35-134	79-35-139	79-35-140	79-35-141	79-35-144	79-35-173
SiO ₂	73.890	73.710	73.550	74.070	73.320	73.900	73.830	72.400	73.450	74.580
TiO ₂	.050	.040	.050	.040	.040	.040	.050	.050	.050	.050
Al ₂ O ₃	14.940	14.450	14.700	14.440	15.090	15.020	14.440	14.930	15.230	14.440
Fe ₂ O ₃	.085	.354	.343	.254	.143	.149	.207	.413	.199	.375
FeO	1.300	1.200	1.500	1.200	1.500	1.000	2.000	2.400	1.000	1.300
MnO	.070	.080	.080	.040	.070	.050	.110	.200	.070	.040
MgO	.130	.130	.140	.120	.120	.120	.170	.140	.120	.130
CaO	.590	.440	.610	.510	.410	.510	.450	.540	.480	.590
Na ₂ O	4.220	3.580	3.890	3.940	4.130	4.140	2.880	3.320	4.210	4.440
K ₂ O	3.890	3.810	3.210	3.410	3.400	4.430	4.110	4.440	3.750	3.380
M ₂ O	.400	.300	.400	.300	.300	.300	.700	.800	.400	.300
M ₂ O	.100	.100	.100	.200	.100	.100	.100	.100	.100	.100
P ₂ O ₅	.470	.480	.440	.420	.470	.430	.440	.500	.530	.440
CO ₂	.100	.100	.100	.100	.100	.100	.100	.100	.100	.100
TOTAL OXIDES	100.053	99.374	99.113	99.504	99.413	100.289	99.487	99.703	99.989	100.215
NORMATIVE MINERALOGY (PERCENT WEIGHT)										
QUARTZ	33.140	37.890	38.000	34.740	35.470	32.110	43.140	40.450	34.110	34.740
CORDIERITE	3.400	4.490	4.890	4.300	4.470	3.420	4.170	4.410	4.270	3.420
ORTHOCASE	23.120	22.780	19.200	21.500	20.310	24.240	21.440	24.530	22.300	22.000
ALBITE	35.790	30.530	33.380	33.740	35.210	34.990	20.340	19.770	35.810	34.400
ANORTHITE	.440	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ACRITITE	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ENSTATITE	.300	.240	.200	.240	.240	.270	.240	.280	.210	.270
FERROSILITE	2.340	1.570	2.280	1.430	2.310	1.510	3.380	4.040	1.180	1.430
MAGNETITE	.120	.810	.500	.370	.230	.700	.590	.900	.290	.390
ILMENITE	.100	.080	.100	.080	.080	.100	.080	.100	.080	.100
APATITE	1.100	1.130	1.010	.940	1.100	.980	1.010	1.170	1.240	1.000
PRASE	0.000	.040	.100	.190	.100	0.000	0.000	0.000	.040	0.000
Na ₂ SiO ₃	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Na ₂ SiO ₃ Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
CaSiO ₃	.230	.230	.230	.230	.230	.230	.230	.230	.230	.230
TRACE ELEMENTS (PPM)										
Ag	.20	.40	.80	1.40	.80	.20	.20	.20	1.00	.20
Ba	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00	20.00
Ca	20.00	30.00	20.00	40.00	20.00	20.00	30.00	40.00	30.00	20.00
Cu	92.00	240.00	500.00	970.00	520.00	44.00	21.00	72.00	290.00	48.00
FLUORINE	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00
Li	780.00	800.00	470.00	590.00	790.00	470.00	1100.00	740.00	440.00	750.00
Mo	2.00	2.00	1.50	2.00	2.00	3.50	2.00	1.00	2.00	3.50
Na	18.00	12.00	10.00	12.00	12.00	8.00	8.00	8.00	18.00	10.00
Si	320.00	990.00	930.00	880.00	870.00	940.00	1340.00	1400.00	850.00	900.00
Sn	330.00	1200.00	2400.00	1500.00	300.00	140.00	420.00	140.00	100.00	100.00
Sr	20.00	10.00	20.00	10.00	20.00	20.00	10.00	20.00	30.00	10.00
Th	4.00	3.00	5.00	3.00	5.00	5.00	5.00	5.00	5.00	4.00
U	33.00	24.00	24.00	25.00	24.00	24.00	27.00	24.00	25.00	25.00
W	23.00	20.00	23.00	14.00	40.00	31.00	14.00	25.00	30.00	14.00
Zn	110.00	99.00	130.00	87.00	190.00	180.00	150.00	190.00	180.00	110.00
Zr	20.00	10.00	10.00	10.00	30.00	-10.00	10.00	-10.00	20.00	10.00

SAMPLE NO	79-35-183	79-35-200	79-35-207	79-43-840	79-43-56	79-43-62	79-43-71	79-43-92	79-43-98	79-43-130	79-43-199	79-43-223
SI02 2	73.750	73.890	70.710	73.730	71.860	73.870	72.120	71.990	75.490	74.390	74.040	77.060
TI02 3	.050	.040	.040	.040	.050	.060	.050	.050	.050	.050	.040	.040
AL203 4	15.180	14.780	15.820	14.400	14.170	14.420	14.910	15.200	14.070	15.080	14.080	12.880
FE2O3 5	.190	.295	.612	.453	.737	.493	.518	.498	.379	.331	.345	.540
FE0 6	.900	1.300	2.500	.600	4.700	2.400	2.800	2.800	1.000	.800	1.300	1.800
MNO 7	.020	.040	.130	.040	.090	.080	.150	.110	.090	.050	.110	.180
MGO 8	.130	.120	.140	.130	.130	.180	.150	.180	.140	.130	.180	.170
CAO 9	.580	.660	.450	1.430	.590	1.270	.740	.420	.420	.590	.460	.470
MA20 12	4.430	3.820	4.030	3.300	3.210	3.300	3.740	.510	3.860	4.510	3.700	1.900
K2O 11	4.320	4.010	3.840	.310	3.050	2.600	4.330	4.430	3.770	3.400	3.920	3.440
M2O 13	.700	.500	.700	.300	.500	.600	.700	.900	.400	.300	.300	.500
M2O- 14	.100	.100	.100	.100	.100	.100	.100	.100	.100	.100	.100	.100
P2O5 15	.450	.440	.350	.480	.340	.470	.470	.420	.130	.430	.130	.200
CO2 14	.100	.100	.100	.300	.200	.200	.200	.400	.300	.200	.100	.200
TOTAL OXIDES	100.500	100.095	99.542	94.813	94.727	97.213	97.978	98.608	100.269	100.361	98.845	99.430
NORMATIVE MINERALOGY (PERCENT WEIGHT)												
QUARTZ	30.770	35.480	30.930	74.100	57.450	41.460	50.440	50.710	37.960	34.740	34.470	51.360
CORUNDUM	3.520	4.190	5.090	16.130	10.970	10.430	9.290	9.590	3.870	3.990	3.480	5.490
ORTHOCLASE	25.480	23.800	23.020	1.940	18.700	15.840	26.370	28.040	22.310	20.140	23.570	20.470
ALBITE	37.440	32.420	34.550	2.880	1.760	2.620	4.360	4.430	32.480	38.120	31.820	16.190
ANORTHITE	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ACMITE	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ENSTATITE	.250	.290	.380	.320	.300	.440	.340	.350	.300	.200	.430	.400
FERROSILITE	1.180	2.080	4.030	.150	7.880	3.380	4.720	3.800	1.440	.800	2.190	3.100
MAGNETITE	.280	.420	.900	.990	1.120	1.030	.780	1.040	.350	.480	.530	.790
ILMENITE	.100	.080	.080	.080	.100	.120	.100	.100	.120	.100	.120	.120
APATITE	1.020	1.000	.800	1.440	.800	1.010	1.120	.980	.280	.980	.280	.280
PYRITE	.020	0.000	0.000	.440	.260	.640	0.000	.040	0.000	0.000	0.000	0.000
SODIUM	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
METASILICATE	.230	.230	.230	.720	.470	.470	.470	.930	.880	.460	.230	.460
CALCITE												
TRACE ELEMENTS (PPM)												
AS 18	.40	.20	.20	5.20	1.80	4.00	.20	.80	.20	.20	.20	.20
BORON 20	20.00	20.00	20.00	10.00	20.00	20.00	20.00	20.00	10.00	10.00	10.00	10.00
BA 21	40.00	40.00	40.00	70.00	60.00	70.00	100.00	90.00	60.00	70.00	60.00	60.00
CU 25	110.00	10.00	31.00	2100.00	1200.00	3000.00	54.00	240.00	9.00	50.00	21.00	39.00
FLUORINE 26	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00
LI 27	490.00	790.00	920.00	100.00	980.00	820.00	500.00	1400.00	1200.00	730.00	1200.00	1100.00
MO 29	2.00	2.00	1.00	7.00	.50	4.50	1.00	.50	4.00	3.50	4.00	1.50
PB 31	8.00	8.00	6.00	18.00	4.00	8.00	4.00	18.00	8.00	4.00	6.00	4.00
RB 32	1070.00	1150.00	1470.00	90.00	990.00	740.00	1470.00	1540.00	1070.00	820.00	1020.00	1290.00
SM 33	180.00	30.00	140.00	8200.00	3900.00	900.00	2400.00	200.00	25.00	50.00	220.00	540.00
SR 34	10.00	20.00	-10.00	30.00	-10.00	10.00	10.00	10.00	-10.00	40.00	-10.00	-10.00
TH 36	5.00	4.00	5.00	4.00	4.00	4.00	4.00	3.00	13.00	4.00	15.00	14.00
U 38	23.00	27.00	23.00	25.00	25.00	27.00	27.00	27.00	27.00	30.00	28.00	27.00
W 40	32.00	25.00	14.00	25.00	33.00	16.00	24.00	21.00	78.00	100.00	12.00	78.00
ZN 42	150.00	110.00	90.00	5500.00	160.00	2800.00	260.00	200.00	49.00	78.00	130.00	89.00
ZR 43	-10.00	-10.00	10.00	10.00	20.00	10.00	20.00	10.00	20.00	10.00	20.00	30.00

SAMPLE NO 49-20-79- 79-35-4.0 79-35-8.4 79-35-20. 79-35-24. 79-35-32. 79-35-34. 79-35-40. 79-35-48. 79-35-50. 79-35-54. 79-35-81.

SI02 2	46.220	57.470	49.800	72.370	74.220	74.170	71.810	73.680	72.520	74.240	73.130	74.030
TI02 3	.080	1.010	.440	.050	.050	.040	.040	.040	.050	.050	.050	.040
AL203 4	17.250	20.450	14.340	14.710	13.410	15.190	14.040	14.430	14.540	14.990	15.000	14.700
FE203 5	.954	.989	.435	.484	.424	.429	.478	.417	.244	.004	1.219	.070
FE0 6	3.200	5.400	4.000	3.000	2.300	1.800	1.900	2.000	2.100	1.200	1.000	1.800
MM0 7	.220	.150	.120	.190	.180	.080	.160	.110	.120	.050	.110	.090
MM0 8	.260	2.590	1.480	.220	.170	.150	.140	.140	.140	.120	.140	.140
CA0 9	1.320	.500	.820	.540	.480	.490	.700	.640	1.060	.460	.460	.460
MA20 12	1.350	2.040	3.190	.730	1.080	5.010	2.940	2.300	3.350	4.220	3.290	3.710
K20 11	4.920	6.000	3.250	5.070	3.390	2.680	3.410	4.150	3.000	4.010	3.500	3.340
M204 13	1.200	1.400	1.000	1.000	.700	.700	.800	.700	.600	.300	.500	.500
M20- 14	.100	.100	.100	.100	.200	.100	.100	.100	.200	.100	.200	.100
P205 15	.840	.220	.140	.370	.200	.140	.500	.430	.420	.400	.470	.470
CO2 16	.100	.300	.200	.100	.100	.100	.100	.100	.100	.100	.100	.100
TOTAL OXIDES	97.734	98.619	99.955	98.734	97.304	100.399	99.158	99.037	98.264	100.034	99.539	99.630

NORMATIVE MINERALOGY (PERCENT WEIGHT)

QUARTZ	37.740	15.880	33.280	44.710	54.380	33.860	39.530	42.950	39.380	31.080	39.910	37.280
CORDIERITE	9.450	10.910	4.990	7.880	7.970	4.010	7.140	5.880	4.740	3.510	5.700	4.540
ORTHOCASE	10.090	14.520	19.450	30.670	20.780	15.890	21.490	24.990	18.200	23.730	21.850	21.130
ALBITE	11.830	17.770	27.230	4.330	9.380	42.450	25.420	19.730	28.980	35.720	28.170	31.530
ANDRINITE	1.110	0.000	1.780	.930	1.390	0.000	.800	1.080	3.290	.430	0.000	.740
ACRITE	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ENSTATITE	.450	4.070	4.210	.540	.410	.340	.330	.410	.330	.300	.350	.330
FERROBILITE	5.200	7.750	4.050	5.500	3.880	1.470	3.370	3.510	3.420	2.180	3.620	3.320
MAGNETITE	1.430	1.480	.950	.710	.430	.470	.710	.420	.340	0.000	1.780	.100
ILMENITE	.140	1.940	1.230	.100	.100	.080	.080	.080	.100	.100	.100	.080
APATITE	2.040	.500	.380	.840	.940	1.000	1.180	.990	.980	.930	1.100	1.100
PYRITE	.340	.020	0.000	.020	.340	.020	0.000	0.000	.250	.040	0.000	.020
LIQUID	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
NETALICATE	.240	.700	.440	.230	.240	.230	.230	.230	.230	.230	.230	.230
CALCITE												

TRACE ELEMENTS (PPM)

AS 18	2.80	.20	.20	.80	2.20	.20	.20	.20	.20	.40	.20	.20
BA 20	30.00	440.00	90.00	20.00	20.00	20.00	30.00	20.00	30.00	20.00	20.00	20.00
BA 21	50.00	1010.00	500.00	80.00	130.00	40.00	50.00	50.00	40.00	50.00	40.00	20.00
CU 25	1400.00	130.00	94.00	190.00	1700.00	110.00	10.00	25.00	1200.00	280.00	53.00	110.00
FLUORINE24	999.00	8800.00	4700.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00	999.00
LI 27	2100.00	390.00	270.00	980.00	960.00	420.00	870.00	1000.00	750.00	880.00	960.00	840.00
MO 29	.50	.50	.50	1.50	1.50	1.50	1.00	1.00	6.00	1.50	3.00	1.50
PB 31	8.00	14.00	14.00	10.00	10.00	12.00	10.00	10.00	12.00	12.00	10.00	10.00
SB 32	1530.00	790.00	550.00	1550.00	1110.00	710.00	1140.00	1320.00	940.00	920.00	1180.00	1040.00
SM 33	200.00	120.00	120.00	540.00	5300.00	80.00	140.00	420.00	1100.00	280.00	400.00	340.00
SR 34	90.00	170.00	120.00	30.00	30.00	40.00	50.00	-10.00	40.00	40.00	10.00	10.00
TM 36	8.00	13.00	8.00	5.00	5.00	4.00	4.00	5.00	4.00	5.00	5.00	5.00
U 38	29.00	3.20	2.10	18.00	30.00	27.00	24.00	25.00	25.00	24.00	24.00	24.00
U 40	37.00	7.00	8.00	15.00	20.00	9.00	21.00	13.00	17.00	17.00	17.00	17.00
ZN 42	240.00	500.00	140.00	700.00	3200.00	340.00	190.00	140.00	3200.00	100.00	250.00	140.00
ZR 43	20.00	180.00	170.00	15.00	-10.00	-10.00	20.00	20.00	10.00	20.00	20.00	10.00

SAMPLE NO	79-43-249	79-43-273	79-43-280	79-43-285	79-43-286	79-43-300	80-35-10
SI02 2	75.240	82.010	75.920	76.100	75.740	75.560	76.670
TI02 3	.060	.050	.050	.060	.040	.040	.500
AL203 4	14.200	11.450	13.470	12.800	13.460	12.420	8.100
FE2O3 5	.300	.270	.221	.308	.125	.590	1.380
FE0 4	.900	.900	.800	1.100	1.300	1.800	3.500
MNO 7	.090	.080	.070	.120	.080	.190	.090
MNO 8	.160	.120	.150	.360	.120	.350	1.110
CAO 9	.430	.350	.410	1.310	.610	1.270	.850
NA2O 12	4.080	.490	4.100	3.620	3.730	1.820	1.650
K2O 11	4.340	1.210	4.140	2.990	3.610	3.420	1.760
M2O+ 13	.300	.300	.300	.400	.300	.600	.800
M2O- 14	.100	.100	.100	.100	.100	.100	.500
P2O5 15	.130	.120	.110	.110	.110	.120	.120
CO2 16	.200	.200	.200	.200	.100	.200	.200
TOTAL OXIDES	100.350	97.850	100.241	99.578	99.345	98.500	97.230
NORMATIVE MINERALOGY (PERCENT WEIGHT)							
QUARTZ	33.960	75.710	35.510	40.450	38.590	49.220	58.180
CORDIERITE	2.740	9.780	2.430	1.940	2.830	4.240	2.830
ORTHOCASE	25.650	7.350	24.480	17.800	21.540	28.690	10.790
ALBITE	34.490	4.240	34.670	30.930	31.920	15.750	14.460
ANORTHITE	.030	0.000	.120	4.630	1.710	4.350	2.240
ANITE	0.000	0.000	0.000	0.000	.350	0.000	0.000
ENSTATITE	.400	.280	.370	.880	.460	.870	2.850
FERROSILITE	1.440	1.380	1.320	1.920	2.450	3.120	4.610
MAGNETITE	.440	.490	.320	.440	0.000	.880	2.090
ILMENITE	.110	.100	.100	.120	.120	.120	.990
APATITE	.280	.290	.230	.230	.240	.290	.290
PYRITE	0.000	0.000	0.000	0.000	0.000	0.000	.190
SODIUM	0.000	0.000	0.000	0.000	.090	0.000	0.000
METASILICATE	.450	.470	.440	.440	.230	.470	.470
CALCITE							
TRACE ELEMENTS (PPM)							
AG 18	.20	.60	.20	.20	.20	.20	1.60
BAROM 20	10.00	-10.00	-10.00	10.00	-10.00	20.00	20.00
BA 21	40.00	50.00	40.00	50.00	80.00	50.00	220.00
CU 23	3.00	40.00	5.50	11.00	4.00	44.00	970.00
FLUORINE 24	9200.00	999.00	999.00	999.00	999.00	999.00	5000.00
LI 27	950.00	1100.00	900.00	1300.00	940.00	1500.00	210.00
MO 29	2.50	320.00	2.50	3.00	3.50	1.50	.50
PS 31	8.00	130.00	8.00	8.00	8.00	6.00	14.00
RB 32	1020.00	450.00	950.00	940.00	1030.00	1290.00	380.00
SE 33	25.00	30.00	17.00	25.00	15.00	540.00	100.00
SM 34	-10.00	-10.00	-10.00	-10.00	-10.00	2140.00	930.00
TH 36	15.00	15.00	13.00	15.00	14.00	15.00	8.00
U 38	29.00	28.00	29.00	28.00	31.00	27.00	1.70
U 40	190.00	1100.00	42.00	100.00	77.00	37.00	13.00
ZN 42	55.00	94.00	40.00	140.00	44.00	110.00	1900.00
ZR 43	20.00	30.00	30.00	40.00	30.00	30.00	250.00

Figure 9-6. Statistical summary of geochemical data for the granitoid rocks from Davis Lake, southwestern Nova Scotia. $\bar{x}$ = arithmetic mean; v = variance; CV% coefficient of variation; Sk = skewness; S = standard deviation. Major elements in weight per cent, trace elements in ppm

	$\bar{x}$	σ	V	Cv (%)	Ku	Sk	95% CI
SiO ₂ (%)	75.83	2.43	5.90	3	7.54	-1.61	73.30-76.40
TiO ₂	0.13	0.12	0.01	90	71.50	8.03	0.11-0.15
Al ₂ O ₃	12.65	1.37	1.88	11	11.43	2.25	12.37-12.93
Fe ₂ O ₃ (T)	1.72	0.53	0.28	31	24.30	4.00	1.62-1.83
FeO	1.21	0.31	0.10	26	7.60	1.61	1.14-1.27
MnO	0.04	0.05	0.003	150	78.90	8.60	0.025-0.048
MgO	0.18	0.11	0.01	62	12.58	3.08	0.16-0.20
CaO	0.67	0.27	0.08	41	8.80	2.12	0.62-0.73
K ₂ O	4.08	0.82	0.67	20	2.15	-1.10	3.92-4.25
Na ₂ O	3.41	0.83	0.69	24	8.05	0.95	3.24-3.58
H ₂ O ⁺	0.76	0.23	0.05	31	8.52	2.08	0.71-0.81
H ₂ O ⁻	0.02	0.02	0.001	140	1.15	1.36	0.01-0.02
P ₂ O ₅	0.11	0.03	0.001	26	-0.61	-0.10	0.10-0.12
CO ₂	0.12	1.42	0.02	116	45.00	6.09	0.09-0.15
Ba (ppm)	73	60.17	3 621	82	9.98	2.29	60-85
Be	14	7.10	50	52	29.49	4.52	12-15
Cu	7	26.00	690	405	62.70	7.58	1-11
F	3318	1042	108 665	31	0.26	0.17	3105-3532
Li	146	64	4104	43	-0.51	0.10	133-159
Mo	2	0.05	0.31	30	1.96	0.33	1.72-1.95
Pb	7	4	18.0	61	2.60	1.30	6-8
Rb	700	122	14 973	17	5.30	0.20	675-725
Sa	50	60	3 715	120	12.40	3.20	37-63
Sr	31	112	153	40	5.58	1.67	28-33
Th	25	6	37	24	2.60	1.20	23-26
U	22	4	14	17	1.90	0.5*	22-24
W	11	4	16	35	0.14	-0.17	10-12
Zn	43	14	204	32	1.70	1.20	40-46

Table 9-7. Statistical summary of geochemical data for the East Kemptville quartz-topaz greisen, tin domain, southwestern Nova Scotia. N = number of samples; $\bar{x}$ = arithmetic mean, mean values for major oxides in weight per cent and for trace elements in parts per million; v = variance; cv% = coefficient of variation; $\delta/\bar{x} \times 100\%$; Ku = Kurtosis; Sk = Skewness; δ = standard deviation; 95% C.I. = 95% confidence limit of mean value; R = range.

N=41	$\bar{x}$	v	cv%	Ku	Sk	δ	95% C.I.	R
SiO ₂	73.50	10.324	4.371	16.3	-2.7	3.213	72.490 - 74.519	57.470 - 82.010
TiO ₂	0.08	0.030	202.066	22.5	4.7	0.027	0.03 - 0.111	0.040 - 1.010
Al ₂ O ₃	14.67	1.734	8.970	9.0	1.6	1.317	14.263 - 15.094	11.650 - 20.450
Fe ₂ O ₃	0.40	0.068	64.150	1.4	0.7	0.260	0.323 - 0.487	-0.125 - 1.219
FeO	1.80	1.138	59.117	2.9	1.6	1.067	59.117 - 1.468	0.600 - 5.400
MnO	0.10	0.002	44.938	-0.3	0.6	0.045	0.087 - 0.115	0.020 - 0.200
MgO	0.25	0.199	174.671	21.5	4.6	0.446	0.115 - 0.396	0.110 - 2.590
CaO	0.67	0.062	36.899	2.8	1.7	0.249	0.595 - 0.752	0.350 - 1.430
K ₂ O	3.63	0.835	25.129	5.0	-1.1	0.914	3.347 - 3.924	0.30 - 6.00
Na ₂ O	2.92	2.215	50.836	-0.9	-0.7	1.488	2.458 - 3.397	0.210 - 5.030
H ₂ O ⁺	0.51	0.063	48.716	2.4	1.4	0.252	0.438 - 0.597	0.300 - 1.400
H ₂ O ⁻	0.11	0.001	27.370	6.2	2.8	0.030	0.100 - 0.119	0.100 - 0.200
P ₂ O ₅	0.36	0.023	41.511	-0.6	-0.6	0.151	0.316 - 0.411	0.110 - 0.680
CO ₂	0.15	0.006	49.314	1.9	1.4	0.075	0.128 - 0.175	0.100 - 0.400
Ba	85	27385	194.410	26.2	4.9	165.486	32.888 - 137.356	20.000 - 1010.000
Cu ³	351	405714	181.350	7.9	2.7	636.957	150.183 - 552.280	5.000 - 3000.00
F	15090	0.6436E+08	53.166	7.6	2.5	8022.836	12557.924 - 17622.564	6700.000 - 49000.000
Li	916	200179	48.839	11.2	2.4	447.414	774.876 - 1057.319	100.000 - 3000.000
Mo	9	2465	497.729	40.9	6.3	49.652	-5.696 - 25.648	0.500 - 320.000
Pb ⁵	12	364	149.317	38.1	6.0	19.083	6.757 - 18.804	4.000 - 130.000
Rb	1034	96275	29.990	1.4	-0.2	310.283	936.697 - 1132.572	90.000 - 1740.000
Sn ⁶	817	2564556	195.779	12.1	3.3	1601.423	312.504 - 1323.447	15.000 - 8200.000
Sr ⁵	75	109755	441.008	40.6	6.3	331.294	-29.447 - 179.691	10.000 - 2140.000
Th	7	18	59.825	-0.4	1.1	4.275	5.797 - 8.496	3.000 - 16.000
U ⁷	24	31	22.385	11.7	-3.2	5.587	23.195 - 26.722	2.100 - 31.000
W	60	28888	281.562	37.4	6.0	169.967	6.718 - 114.014	7.000 - 1100.000

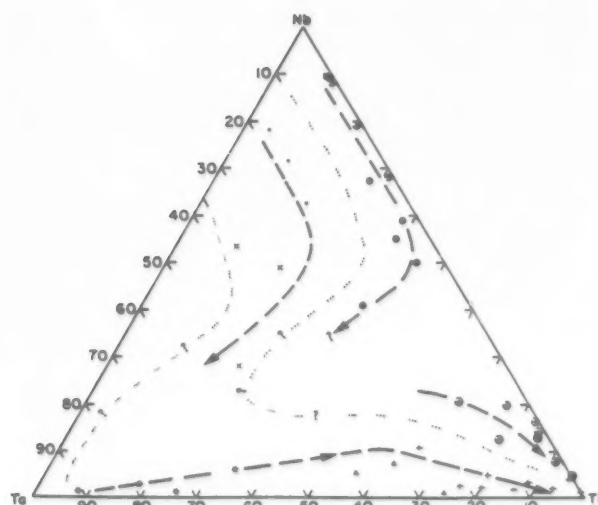


Figure 9-14. Nb-Ta-Ti plot showing the changes in composition of cassiterite from various rock types. dots = porphyritic biotite granite; ringed points = Cambro-Ordovician metasedimentary rocks; crosses = sericitized granite; asterisks = albitized granite; triangles = albitite breccia; plus signs = quartz-topaz greisens; ringed plus signs = veins in metasediments.

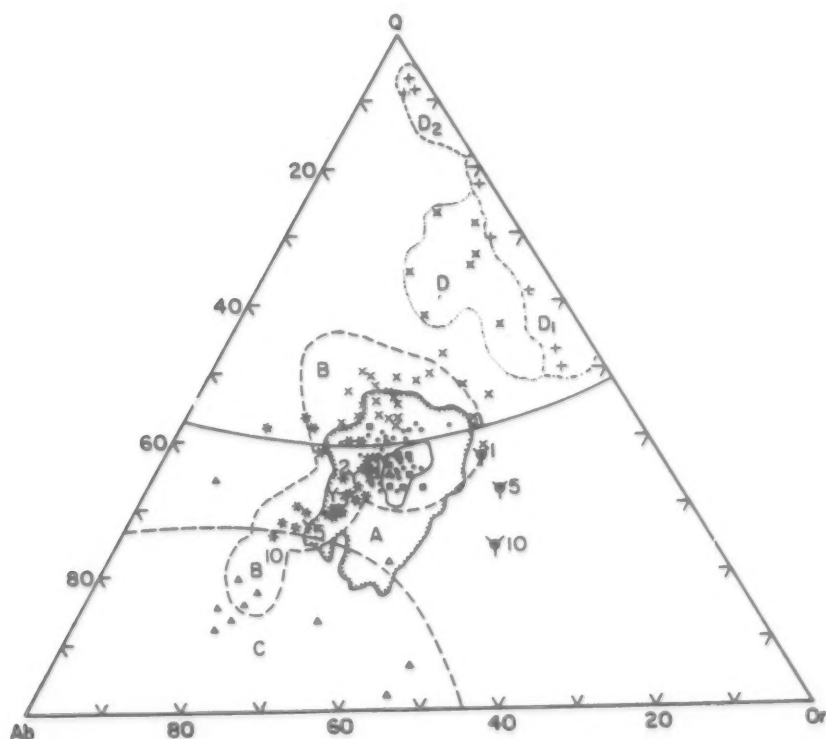


Figure 9-15. Projections of compositions of granitic rocks from Davis Lake area on to CIPW normative Q-Ab-Or ternary. A-area of predominant granite compositions with contours at 90 and 95%; B-area of tin-bearing granites including albitized and li-rich varieties; C-area of feldspathized rocks including metasomatized albitites; D- area of greisenized granites (D1 mica-quartz greisen and D2 quartz-mica greisen). Definition of symbols: Y superimposed on circle- ternary minimum at pressure 1,5 or 10 kb as numbered (PH20 = PT); Y accompanied by figure 2,3,5 or 10, ternary minimum at 2,3,5 or 10 kb 'dry', respectively (PH20=0); unbroken curved line- feldspar-quartz cotectic at 500b; dots- porphyritic biotite granite; crosses(x) superimposed on dots (area D)- greisenized granites; plus signs(+)- greisens; filled triangles- albitites; open triangles- highly altered granites; asterisks- albitized granites; filled squares- average Sn- bearing granites from various provinces.

L.E.	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	K ₂ O	Na ₂ O	H ₂ O ⁺	H ₂ O ⁻	P ₂ O ₅	CO ₂	Be	Ba	Cu	F	Li	Mo	Pb	Rb	Sn	Sr	Th	U	W	Zn	
SiO ₂	1.00																											
Al ₂ O ₃	-.21	1.00																										
Fe ₂ O ₃	-.77*	.20	1.00																									
FeO	-.53*	.20	.19	1.00																								
MnO	-.60*	.02	.22	.73*	1.00																							
MgO	-.61*	.24	.24	.50*	.53*	1.00																						
CaO	-.15	-.11	.17	-.11	.18	-.29	1.00																					
K ₂ O	-.20	-.01	.24	.14	.25	.27	.07	1.00																				
Na ₂ O	.20	.08	.21	-.60*	-.63*	-.69*	-.33	.17	1.00																			
H ₂ O ⁺	-.66*	.12	.21	.42*	.37*	.25	.24	.01	-.16	1.00																		
H ₂ O ⁻	.13	-.02	-.07	.01	-.00	.13	-.05	.02	-.11	-.04	1.00																	
P ₂ O ₅	-.08	-.06	.29	-.10	-.19	.10	.02	.32	-.03	.19	.05	1.00																
CO ₂	-.13	-.16	.13	.06	.05	.31	.01	-.30	-.17	.02	.03	.10	1.00															
Be	-.05	.10	.21	.02	-.04	.08	-.06	-.05	.28	-.05	.30	-.02	.16	1.00														
Ba	-.36*	-.08	.03	.57*	.45*	.34	.61*	.10	-.07	-.56*	.30	-.02	.15	.18	1.00													
Cu	-.23	-.18	.15	.21	.19	.65*	.05	.51*	.32	-.21	.01	-.00	.40*	.21	.06	1.00												
F	-.30	.02	.16	.66*	.51*	.51*	.38*	-.18	.67*	.39*	.03	.07	.16	.13	.11	.22	1.00											
Li	-.65*	.13	.09	.52*	.47*	.32	.33	-.09	.03	-.62*	.38*	-.01	.15	.18	.03	.05	.26	1.00										
Mo	-.04	-.22	-.14	.18	.29	.36*	-.02	.10	.17	-.26	.10	.11	.33	.08	.16	.20	.59*	.64*	1.00									
Pb	-.42*	.08	.41*	.43*	.41*	.48*	.31	.07	.69*	.38*	.13	-.01	.33	.08	.16	.20	.59*	.64*	.20	1.00								
Sn	-.55*	.04	.28	.49*	.38*	.21	.35*	-.12	-.13	-.66*	.51*	-.09	.19	-.07	-.01	.46*	-.04	.11	.37*	.07	.13	1.00						
Sr	.22	-.03	.14	-.67*	-.66*	-.47*	-.40*	.33	-.11	.77*	.27	.01	.32	.12	.18	-.48*	-.05	-.53*	.38*	.11	.29	.49*	.49*	1.00				
Th	-.55*	.20	.16	.65*	.56*	.36*	.65*	.01	-.01	-.44*	.39*	-.11	.13	-.02	-.06	.56*	.20	.29	.62*	.22	.26	.49*	.49*	1.00				
U	-.40*	.21	.30	.31	.24	.33	.33	.01	-.11	-.08	.26	-.04	-.04	-.12	-.20	.21	-.11	.04	.16	.01	.08	.28	.21	.39*	1.00			
W	-.07	.05	-.04	.01	-.02	-.01	.05	.13	.06	.06	-.01	-.06	-.02	.23	-.16	-.01	.05	-.02	.18	-.04	.00	-.01	.08	.13	.11	1.00		
Zn	-.49*	.09	.09	.60*	.54*	.44*	.27	.18	-.06	.53*	.51*	-.01	-.08	.10	-.10	.53*	.28	.19	.53*	.33	.31	.50*	.63*	.65*	.27	.10	1.00	
n	.13	.39*	.06	-.05	-.13	-.16	.04	-.01	-.26	.38*	-.01	.04	.11	.18	-.18	-.25	-.11	-.19	-.13	-.01	-.18	-.23	.30	.04	.14	-.07	-.15	1.00

Figure 9-16. Matrix of correlation and coefficient based on log10 transformed data of 94 samples of granitoid rocks. Values with an asterisk denotes significant correlation at a 99% level of significance. Numerical values for the significance is from Fisher and Yates(1963).

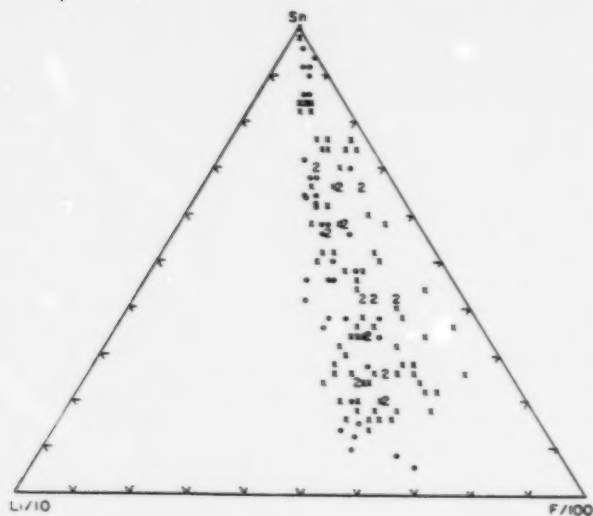


Figure 9-18. Ternary Sn-Li-F diagram showing the trend for the granitoid rocks and quartz-topaz greisen, East Kemptville area.

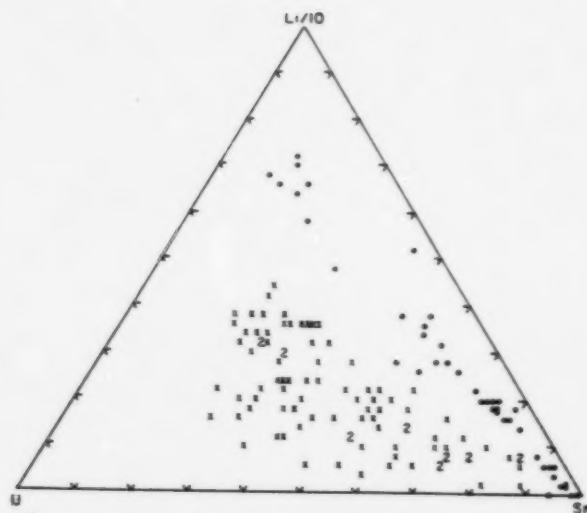


Figure 9-19. Ternary Li-U-Sn diagram showing the trend for the granitoid rocks and the quartz-topaz greisen, East Kemptville area.

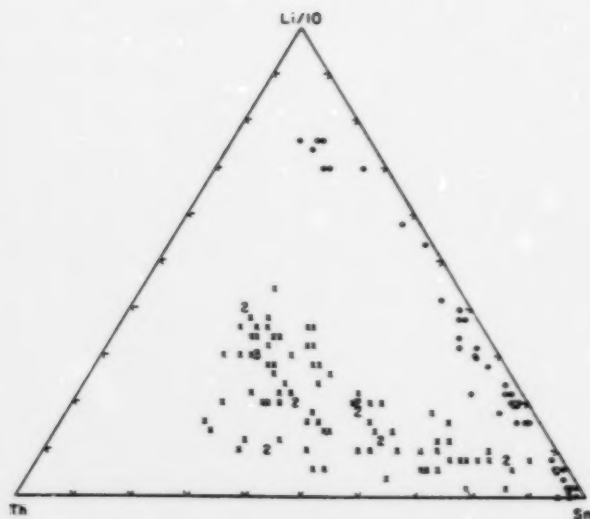


Figure 9-20. Ternary Li-Th-Sn diagram showing relationship in granitoid rocks and quartz-topaz greisen, East Kemptville area.

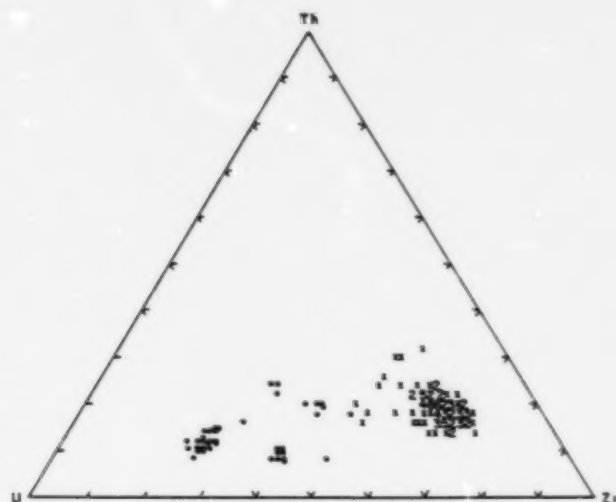


Figure 9-21. Ternary Th-U-Zr showing the separation of granitoid rocks from the greisen, East Kemptville area.

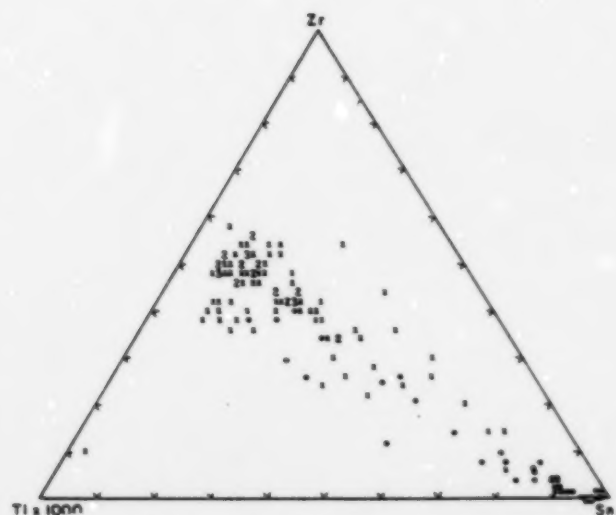


Figure 9-22. Ternary Zr-Ti-Sn showing the relationships in granitoid rock and greisen, East Kemptville area.

Table 9-8. Parameters used to distinguish between the East Kemptville granites and greisens by discriminant function analysis as discussed in the text.

After step 22 variables included						After step 22 variables not included				
Variable	Wilks Lambda	Equivalent F	Variable	Tolerance	F Removed	Wilks Lambda	Variable	Minimum Tolerance	F Remaining	Wilks Lambda
F	0.23112	449.105	TiO ₂	.394	1.52	0.036	SiO ₂	.158	.4591	.0360
Th	0.14373	399.135	MnO	.465	3.96	0.037	Al ₂ O ₃	.159	.3323	.0360
W	0.09977	400.017	MgO	.437	3.04	0.037	Fe ₂ O ₃	.153	.0162	.0361
MgO	0.07873	386.108	CaO	.430	11.48	0.039	FeO	.156	.0159	.0361
Zr	0.06986	348.785	K ₂ O	.190	9.25	0.039	H ₂ O ⁺	.132	.0376	.0361
Na ₂ O	0.06276	323.533	Na ₂ O	.363	52.23	0.052	CO ₂	.160	.7376	.0359
FeO	0.05782	300.278	H ₂ O ⁻	.868	1.22	0.036	Sn	.157	.3378	.0360
Cu	0.05389	280.895	P ₂ O ₅	.352	3.60	0.037	Sr	.160	.1740	.0360
CaO	0.05069	264.229	Ba	.702	2.69	0.036				
Zn	0.04778	251.063	Cu	.483	9.70	0.039				
U	0.04548	238.616	F	.291	12.06	0.039				
Mo	0.04341	227.696	Li	.330	2.14	0.036				
MnO	0.04150	218.504	Mo	.392	6.20	0.038				
FeO	0.04150	238.634	Pb	.676	2.29	0.036				
K ₂ O	0.04050	224.130	Rb	.160	2.28	0.036				
Li	0.03923	213.395	Th	.331	22.33	0.043				
P ₂ O ₅	0.03849	201.484	U	.462	4.86	0.037				
Ba	0.03785	190.628	W	.490	47.53	0.050				
Rb	0.03742	180.042	Zn	.386	4.84	0.037				
Pb	0.03701	170.568	Zr	.321	9.59	0.039				
TiO ₂	0.03652	162.449								
H ₂ O ⁺	0.03614	154.684								
Eigenvalue						Chi-Squared				
22.67						415.04				
% Variance						Wilks Lambda				
100						0.9817				
Canonical Correlation						0.03614				

discrimination. Table 9-8 also shows that FeO was included at Step 7 to achieve significant discrimination but at Step 14 it was removed from the analysis. The resulting standardized discriminant function is as follows:

$$DF = 0.4656 \log CaO + 0.1850 \log TiO_2 - 0.2715 \log MnO - 2.2463 \log MgO + 0.6339 \log K_2O - 0.9417 \log Na_2O + 0.1117 \log H_2O + 0.2977 \log P_2O_5 - 0.1832 \log Ba - 0.4072 \log Cu - 0.5791 \log F - 0.2391 \log Li + 0.3664 \log Mo - 0.1726 \log Pb - 0.3533 \log Rb + 0.7328 \log Th + 0.3001 \log U - 0.7839 \log W - 0.3279 \log Zn + 0.4966 \log Zr.$$

The result of the discriminant function analysis shows the granitoid rocks and the greisens from the Davis Lake complex can be successfully differentiated on the basis of 9 major and 11 trace elements. The derived discriminant score for each, using the above function, is shown in a histogram (Fig. 9-24). From Table 9-8 and Figure 9-24 it is clear the initial aim of separating the two is satisfied and the two are completely separated. However, there are a few significant features of the discriminant analysis which deserve closer attention.

Firstly, the discriminant function which separates the two populations requires a complex arrangement of 20 variables suggesting that many of these elements behave similarly in the two groups of rocks, and this is in agreement with

the discussion in the previous section on the covariance relationship of the elements. Also, CaO, K₂O, Na₂O, Cu, F, Rb, Li, Th, U, W, Lu and Zr are significant discriminators. Various ternary relationships (Figures 9-18 to 9-22) involving many of these elements either show systematic continuous evolution (magmatic stage to greisenization stage) or separate the two populations, or both.

Secondly, the histogram of the discriminant scores (Fig. 9-24) indicates the distribution of the score is polymodal for the granitoid rocks, suggesting that there are various distinct rock types amongst the granites, and this is in agreement with the petrographic descriptions given above. The mean discriminant score for the greisen was -7.844 while for the granitoid rocks it was 3.350 and these are considered as group centroids.

Thirdly, the major element oxides SiO₂, Al₂O₃, Fe₂O₃, FeO and CO₂, and the trace elements Sn and Sr are excluded from the discriminant analysis. This exclusion may be interpreted in terms of a geological meaning, i.e. although the greisen and the granitoid rocks are separated by a complex arrangement of major and trace elements, these elements are not important contributors to this discrimination. This simply indicates the Sn-greisen at East Kemptville is genetically related to silica-rich peraluminous tin-granite of the Davis Lake Pluton, as best shown by these elements. For example, if the

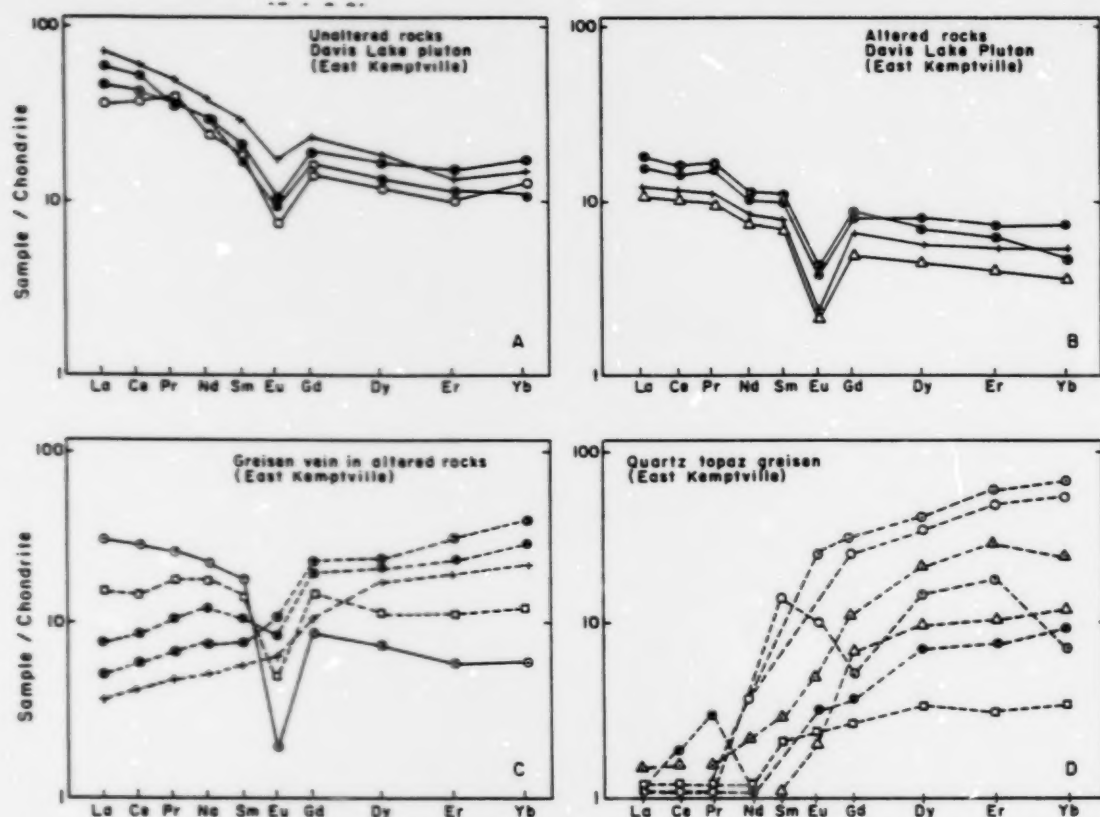


Figure 9-23. Chondrite normalized REE patterns of altered, unaltered and mineralized rocks from the East Kemptville area. For explanation see text.



Figure 9-24. Frequency distribution of discriminant scores for the granitoid rocks and the greisens. All values exceeding 12 are grouped in the last interval.

tin-greisen at East Kemptville was not related to the granitoid rocks, certainly tin should have been one of the important discriminators between the two groups of rocks. The exclusion of CO_2 suggests that its role in the evolution of magmatic rocks leading to greisen formation may be important, a possibility also indicated from the REE data.

CONCLUSIONS

Although reconnaissance in scope, this investigation based on field, petrologic and geochemistry of the granitoid rocks of East Kemptville area shows there is a systematic evolution from the earliest granodiorite to the latest quartz-topaz greisen and this sequence essentially reflects a continuous change from magmatic to hydrothermal processes. The overall mineralogy and geochemistry suggests that the granites are essentially the product of a process controlled by fractional crystallization leading to late stage residual magmas that were geochemically very evolved and specialized in granophile trace elements. This evolution largely involved crystal-melt fractionation with evolution of a volatile phase. The concentration of tin and other metals requires several processes operating within a localized area and, within the limited area, volatile saturation and migration seems to be critical.

CHAPTER 10

EAST KEMPTVILLE TIN PROJECT*

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INTRODUCTION

Rio Algom Limited made a decision to proceed with the development of the East Kemptville tin deposit in October 1983. This decision was the result of intensive evaluation work in the areas of geology and metallurgy as well as careful studies of environmental and marketing aspects. The property is situated about 54 kilometers northeast of Yarmouth, Nova Scotia, near the village of East Kemptville.

The geological setting is a granitic intrusive at the contact with metasediment. The tin mineralization is in the form of cassiterite (SnO_2), which is about 80% tin. Sulfides are associated with the deposit containing silver, copper and zinc, which, however, are recoverable as by-products. The mineralization is associated with quartz veins and greisen alteration. In the center of the deposit, the zones are so close together that they have the appearance of a massive or porphyry-type deposit.

ORE RESERVES

Rio Algom's mine plan is based on 56 million metric tons at 0.165% tin. During the earlier years, up to year nine, feed to the mill will average about 0.2% and lower grade material will be stockpiled. During the later years of production the average grade will be 0.16% tin. A 0.08% cutoff grade is planned for the entire 17 years of production.

East Kemptville is a relatively low grade tin deposit, and only a combination of favorable circumstances have made it economic.

(A) The ore body is situated at, or very close to the surface, which permits mining by open pit methods at a very low stripping ratio.

(B) The tin occurs in relatively large crystals that allow for good recovery by gravity techniques.

(C) It is ideally situated in a well-settled area, with all essential infrastructural requirements already in place.

(D) Rio Algom has been encouraged by the very positive and cooperative response from various governmental agencies involved in the project.

MINE PLAN

The mine plan incorporates an initial pit, followed by three expansions to the ultimate pit. Mining will be at 14 000 metric tons per day, which will include 9 000 tons of ore plus 5 000 tons of waste and low grade. After five years, the mining rate will increase to 21 000 tons per day.

Plans are to drill vertical 6 1/2 in. blast holes on a six by six metre pattern. Bulk slurry explosives will be used to break the rock. Loading will be with hydraulic shovels into 70-ton haulage trucks. The trucks will deliver the ore to a 42 by 65 in gyratory crusher in which it

will be crushed to six inches. Secondary and tertiary cone crushers will be used to further crush the ore to 1/2 in.

PROCESS DESIGN

Tin extraction technology is practically unknown in North America, but Rio Tinto-Zinc Corporation which has a 53% ownership in Rio Algom, is involved in tin production in the United Kingdom, and Rio Algom has been able to utilize this experience and expertise, both in its metallurgical evaluation work, and in tin-related aspects of plant design.

During the past 15 months, Rio Algom has conducted intensive mineralogical and metallurgical test work, both on laboratory and pilot plant scales. The mineralogical work was done largely in the United Kingdom, the laboratory test program was performed at the Lakefield Research Facility in Ontario, and the pilot plant work was conducted in Australia. This program has resulted in significant improvements in recovery and concentrate grades over best projections that could be made before this work was conducted.

CONCENTRATING PROCESS

The concentrator is being designed to process an average of 9 000 tons per day. The first step in the concentrator is to reduce the size of the ore to minus 700 microns in a rod mill/ball mill grinding circuit. The ball mill will operate in closed circuit with high capacity Banana-type screens.

The screened product will first be deslimed to plus 30 microns with cyclones, then passed over the primary gravity separation process with Reichert cones.

Tin in the form of cassiterite has a specific gravity of 6.8. The common method of separating cassiterite from associated gangue is by methods that concentrate the high specific gravity particles. The Reichert cones to be used are high capacity separators. About 65% of the throughput is expected to be rejected in this stage with a loss of only 5% of the contained tin.

Further upgrading of the concentrate will be achieved with spirals, then tables. The spirals are medium capacity gravity separators. They work by carrying the stream of ore down a concave spiral ramp. The heavy particles concentrate on the inside of the ramp.

Tables are the final cleaning step of the gravity process. These are low capacity machines, which, by a combination of shaking in one direction with water washing over a slope in the other direction, are able to make a very selective gravity separation.

As the tin is concentrated in the gravity circuit, sulfides will also be concentrated. They are like tin, much heavier than the waste components of the rock. To further upgrade the tin concentrate, sulfides will be removed by

200 Granites & Mineral Deposits SW N.S.

froth flotation. In this step, sulfides are separated, then in a further cleaning step, copper and zinc concentrates are recovered.

The tin concentrate also contains one other heavy mineral, an aluminum silicate called topaz. To remove the topaz, it will be necessary to dry the concentrate and separate the product in an electrostatic separator. Cassiterite being an electric conductor, is easily charged, then separated on an electrostatic drum from the nonconductive topaz.

In the process, two grades of concentrate will be produced. A relatively coarse fraction containing about one third of the tin will be produced in the gravity circuit at about 65% grade. A second concentrate containing more middling impurities, as well as the remainder of the tin, will grade about 17%. This material will be upgraded to about 40% in the electrostatic process.

Basic design parameters for the plant are: Daily throughput - 9 000 metric tons/day; average head grade - 0.165% Sn; tin recovery - 73%; annual tin production - 4 200 to 4 700 tons of tin in concentrate; mine operation - 2 shifts, 7 days/week; concentrator - 3 shifts, 7 days/week, and number of employees - 250.

SITE LAYOUT

The site layout is relatively compact. The concentrator is situated as near as possible to the mine, with the tailing pond being adjacent to the concentrator. Tailing from the concentrator will be pumped to the tailing area. Because the tailing is coarse, it will be easily deposited onto a free-draining berm. Except for the small amount of flotation reagents, there are no chemicals added to the process. Consequently, the tailing is composed of coarse ground rock about the consistency of beach sand. Water will be recirculated from the tailing back to the plant.

The development schedule is as follows: December 1983 - start detail engineering; April 1984 - start site work; June 1984 - start construction of tailing dams; October 1984 - complete tailing dams; December 1984 - all buildings to be enclosed; July 1985 - start pre-production mining; November 1985 - start-up of concentrator, and January 1986 - mill to be in commercial production.

ENVIRONMENTAL CONCERNS

Before commencing construction, Nova Scotia government approvals in a number of areas will be required. These include water rights for the process water supply, a waste disposal permit for the tailing area, and a mining lease to permit mining of the deposit. These approvals are currently being negotiated.

Environmental protection is a paramount concern in Rio Algom's thinking as well as in government environmental agencies. A carefully planned and effective scheme to insure surface and groundwater quality control, as well as other

concerns such as air quality and general ground disturbance, are basically the objectives that both are seeking. These discussions are proceeding well.

MARKETING

Tin is generally sold by a mine such as East Kemptville in the form of concentrate to a custom tin smelter, which in turn sells refined tin metal to buyers based on prices on the Penang Straits tin market or on the London Metal Exchange. Toll charges for smelting are calculated on the basis of concentrate grade and quality. There is a major excess in tin smelting capacity in the world, so that contract rates are very competitive.

Rio Algom is negotiating with several smelters for a long term smelting contract. As an alternative strategy, it is investigating the possibility of building a smelter in Canada, probably at the East Kemptville site. In this scenario the company would be marketing tin mainly in North America. However the pricing would still be based on the London Metal Exchange.

The price of tin is regulated to the extent possible by the International Tin Council, an association of tin producing and consuming countries. Control is achieved by reducing producer quotas during periods of low demand, and by providing incremental supplies of tin from a buffer stock when demand is high.

Canada is presently a consuming member of the International Tin Council, and at the present time the country consumes slightly less tin per year than East Kemptville will produce.

ECONOMIC IMPACT

The most obvious benefit from the development of this project will, of course, be the 250 jobs created by direct permanent employment, and probably more than double that number of jobs that will be needed in service and support requirements from existing businesses or new ones that will be set up.

Rio Algom expects to initiate a major training program for the specialized operating skills required, but it also believes that a good capability for maintenance requirements will be available within the province. Manpower requirements are expected to be mostly satisfied within Nova Scotia. Adequate housing is believed to exist in the area, and there are no plans for a company-sponsored housing program.

The project will provide a payroll of about C\$8 million per year, and various supply requirements will approximate C\$10 million. With an average power demand of some 10 MW, it will provide an important base load for the Nova Scotia Power Commission. Finally, there is the matter of tax revenue. It will be an important benefit to the area municipalities, and to the province as a whole.

CHAPTER 11

GEOLOGY AND GEOCHEMISTRY OF THE CARBONIFEROUS WEDGEPORT PLUTON
HOSTING Sn-W MINERALIZATION

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Chatterjee, A. K., Strong, D. F., Clarke, D. B. and Keppie J. D., Geology and geochemistry of the Carboniferous Wedgeport Pluton hosting Sn-W mineralization; in Guide to Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

ABSTRACT

The 316 Ma old Wedgeport Pluton is intrusive into deformed and regionally metamorphosed turbidites of the Goldenville Formation. During the Acadian Orogeny, the metasedimentary rocks were deformed by upright, north-south subhorizontal, tight folds associated with cleavage fans, bedding/cleavage intersection and flexural slip lineations. Regional low pressure metamorphism followed the folding producing andalusite-cordierite $\pm$ staurolite $\pm$ garnet-bearing assemblages in lithological units of appropriate compositions.

The Wedgeport Pluton is essentially a biotite monzogranite with a strong airborne radiometric signature. The Pluton shows marked variation in textural, mineralogical and geochemical characteristics, however, when considered in its overall setting, the mineralogical and geochemical evolution is easily explained by considering processes at magmatic (e.g. fraction of plagioclase and biotite) and at postmagmatic (e.g. volatile release, alteration of pre-existing minerals) stages.

The Goose Bay granitic facies occupies the eastern marginal part of the Wedgeport Pluton and the major constituents are faintly perthitic alkali feldspar, oligoclase, biotite, quartz and muscovite. The Pinkney Point granitic facies includes many petrographically distinct rock types showing pronounced effects of hydrothermal alterations and attending mineralization. The endo-contact mineralization is characterized by impregnation and shear-related veins of greisen in the greisenized granite. The mineral associations are complex and probably represent different episodes of mineralization. The ore minerals are cassiterite, wolframite, scheelite, molybdenite and silver-bearing sulphosalts together with pyrite, pyrrhotite, arsenopyrite, native bismuth and rarely an uranium molybdate (sedorite?).

STRUCTURAL SETTING OF THE WEDGEPORT PLUTON

The geology and structural setting of the Wedgeport Pluton has been illustrated in Figure 11-1. It is clear from the map that the metasedimentary rocks of the Cambro-Ordovician Meguma Group occur predominantly in two stratigraphic horizons. These areas are on the Tusket Wedge and along the Little River. The exposures on the Tusket Wedge lie close to the core of a major anticline and are thus the stratigraphically lowest unit exposed in the area. The unit is made up of a series of finely banded pelites and psammites exhibiting Bouma units B and E, and B, C and E, and may have been deposited in proximal levee, back-levee or distal turbidite fan environments (Walker, 1979). The Little River horizon consists of massive rusty greywacke and related pelitic schists, and often display the complete classical Bouma Sequence suggesting that it was deposited in the outer suprafan lobes or outer fan environment. Irregular lenses and nodules of calc-silicate rocks occur throughout the area. Most appear to be secondary in origin and developed at various times in the geological history of the area.

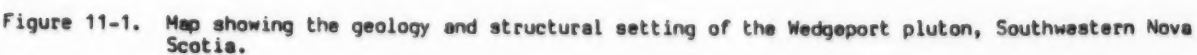
The earliest structure observed in the rocks is a foliation parallel to the bedding planes which is defined by the parallel alignment of chlorite, muscovite and biotite. In the absence of a fold phase related to this foliation, it is inferred that this foliation is produced by loading.

The rocks west of the Tusket River lie on the west limb of a major north-south trending anticline. Minor structures with this phase of folding include cleavage, an intersection lineation produced by the intersection of bedding and cleavage, an "a" lineation produced by bedding plane slip and a few minor folds (F1). These minor folds are generally tight, upright, north-south trending folds with gently plunging fold axes (Fig. 11-2a). The steeply dipping overturned attitudes of the bedding throughout the area (Fig. 11-1) reflect these observations.

The S1 cleavage is present at most outcrops and it ranges from a fracture cleavage in the psammitic layers to a slaty cleavage in the pelitic units. This cleavage is defined by parallel alignment of chlorite, muscovite and biotite, and it is usually oriented north-south and steeply dipping (Fig. 11-2a). Most of the variation in cleavage attitude is due to cleavage refraction and perhaps the minor dispersion of cleavage poles around the periphery of the plot is due to later deformation. The S1 slaty cleavage appears to be parallel to the F1 fold axial plane.

The intersection between bedding and S1 cleavage produces the L1 intersection lineation which is invariably present at most outcrops. This lineation is gently plunging to the north or south and appears to be roughly parallel to the mean F1 fold axes deduced from the π -circles through the bedding plane poles (Fig. 11-2a). A moderately dipping crenulation cleavage (Fig. 11-2b), which is difficult to observe in the field, is invariably present in thin section and it deforms both the bedding and the S1 cleavage. The orientation of the crenulation axes ranges considerably because these were superimposed upon already folded bedding planes and F1 cleavage fans.

The most prominent set of joints in the area are north-easterly trending and steeply dipping to the southeast (Fig. 11-2c). These joints cut both the rocks of the Goldenville Formation and the granitic rocks of the Wedgeport Pluton. The joints are often associated with tension gashes which formed at an oblique angle to the master joints. Measurements of these associated features allow the causative stress directions to be deduced. It is clear that the intermediate stress axes display a systematic variation but all lie in the mean shear joint plane (Fig. 11-2c). The minimum stress axes, σ_3 , appears to vary in a small circle at about 30° from the mean shear plane pole, whereas the maximum stress axes, σ_1 , vary about a 60° small circle. The deduced sinistral sense of movement on these master joints agrees with observed displacements. The sinistral shear joints are parallel to several large faults in the area (Fig. 11-1) and probably these northeast-southwest shears are



related to the large-scale northeastward movement of Acadia relative to cratonic North America during the early Carboniferous (Kent and Opdyke, 1978).

A set of dextral cross joints displaces the northeast-southwest joints (Fig. 11-2d), and tension gashes associated with these suggest that the maximum stress is northwest-southeast and subhorizontal, and the minimum stress is west-southwest/east-northeast and subhorizontal. Thus, this set of cross joints are not genetically related to the earlier northeast-southwest joints. However, in several instances, greisen developed along these cross joints is similar to that associated with the northeast-southwest joints. This might suggest that they were not separated by a great time interval. A large number of felsic dykes and quartz-tourmaline veins occur along the coastline at Pinkney Point. They have a similar orientation (Fig. 11-2e) suggesting they were intruded under the same stress regime. Thus, they are probably late phases of the Wedgeport Pluton.

Within the map area there are three large northeasterly trending faults which cut the Wedgeport Pluton and form its northwestern margin (Fig. 11-1). These faults are parallel to the northeast-southwest sinistral shear joints described earlier and presumably genetically connected. In the northern part of the area there are three east-northeast/west-southwest faults which truncate the northeasterly trending, magnetic marker horizons in the Goldenville Formation. The largest of these is the most northern fault zone running through the Dominique polymetallic tin prospect (DDH 77-25, 77-32). Intense shearing associated with these faults has produced the graphite schist. Complex fracture-filling Sn-Cu-Zn-Bi-Ag-W mineralization is closely associated with the chloritization of the wall rocks.

PETROGRAPHY OF THE GRANITIC ROCKS

The granitic rocks of the Wedgeport Pluton have been divided into three groups: (a) unaltered biotite monzogranite, (b) 'paraintrusive' granitic rocks and (c) mineralized granitic rocks. The unaltered monzogranite is designated as Goose Bay facies and the remaining two groups were mapped as Pinkney Point facies of the Wedgeport Pluton. The boundary between the two facies of the granitic rocks is not exposed but is inferred to lie along the magnetic, north-south trending, mafic dykes which are presumed to have intruded along a weak zone. The northern boundary of the unaltered monzogranite is inferred to be the magnetically anomalous, northeast-southwest fault zone intersected in DDH 78-33. The northwestern margin of the 'paraintrusive' and mineralized granitic rocks is also a northeast-southwest fault zone which was the focus of intense hydrothermal alteration, greisenization and mineralization observed in DDH 80-3, 80-2 and 80-1 (See Fig. 11-1).

UNALTERED BIOTITE MONZOGRANITE

The unaltered biotite monzogranite is grey, homogeneous, equigranular, coarse grained, massive

rock. The average grain size is eight millimetres and the texture is hypidiomorphic granular. The major constituents are alkali feldspar, quartz, plagioclase, biotite, and common are epidote, garnet, ilmenite, apatite, zircon and very fine grained muscovite. The grain boundaries between the feldspar minerals are usually sutured whereas the quartz-feldspar boundaries are usually more curved. The quartz grains usually show an undulatory extinction and, in a few cases, are granulated. Quartz occurs partly as euhedral or subhedral grains but often it is anhedral and fills the interstices between other minerals. A small percentage of the quartz occurs as concave or drop-like inclusions in alkali feldspar, as a micrographic intergrowth with alkali feldspar, and rarely, plagioclase, and as a myrmekitic intergrowth with plagioclase. The alkali feldspar usually occurs as subhedral or anhedral grains and frequently it fills the interstices between plagioclase laths and subhedral or euhedral quartz grains. The biotite is usually euhedral against plagioclase and sometimes it occurs as small flakes in feldspars. It occurs usually as independent flakes or as an intergrowth with muscovite. The biotite is usually a green or brownish green color but occasionally a dark reddish brown variety is also found in samples. The accessories garnet, ilmenite, apatite and zircon occur as euhedral to subhedral inclusions in biotite. Garnet has been found as inclusions in plagioclase feldspar and it may also show grain boundaries against biotite and K-feldspar.

In the unaltered biotite monzogranite, quartz and feldspar form two types of intergrowth, i.e. micrographic and myrmekite, and these are briefly described because they may provide information on the reaction sequences in the granite. The micrographic intergrowth is usually the radiating fringe or insular type, but vermicular and runiform types also occur. In the radiating fringe type, the core is usually alkali feldspar and rarely plagioclase. In some alkali feldspar grains, there are commonly several domains that differ from each other in the orientation of the quartz grains. Usually the orientation of the quartz changes abruptly at the grain boundaries, but often continues with the same optical orientation from one alkali feldspar grain into another. Rarely, the micrographic quartz continues as vein-like bodies between two feldspar grains, or as the intergranular quartz along the cleavage planes of the alkali feldspar grains.

Myrmekite sodic plagioclase usually forms rims between plagioclase and alkali feldspar, and the following characteristics are common:

(a) the boundary between plagioclase and the myrmekite rim is commonly smooth whereas the boundary between the rim and alkali feldspar tends to be irregular. Besides the common appearance of vermicular quartz, the rim-plagioclase boundary is often masked by a sharp change in the plagioclase composition;

(b) the rim plagioclase is generally in crystallographic continuity with the host plagioclase and the twinning of the host plagioclase either

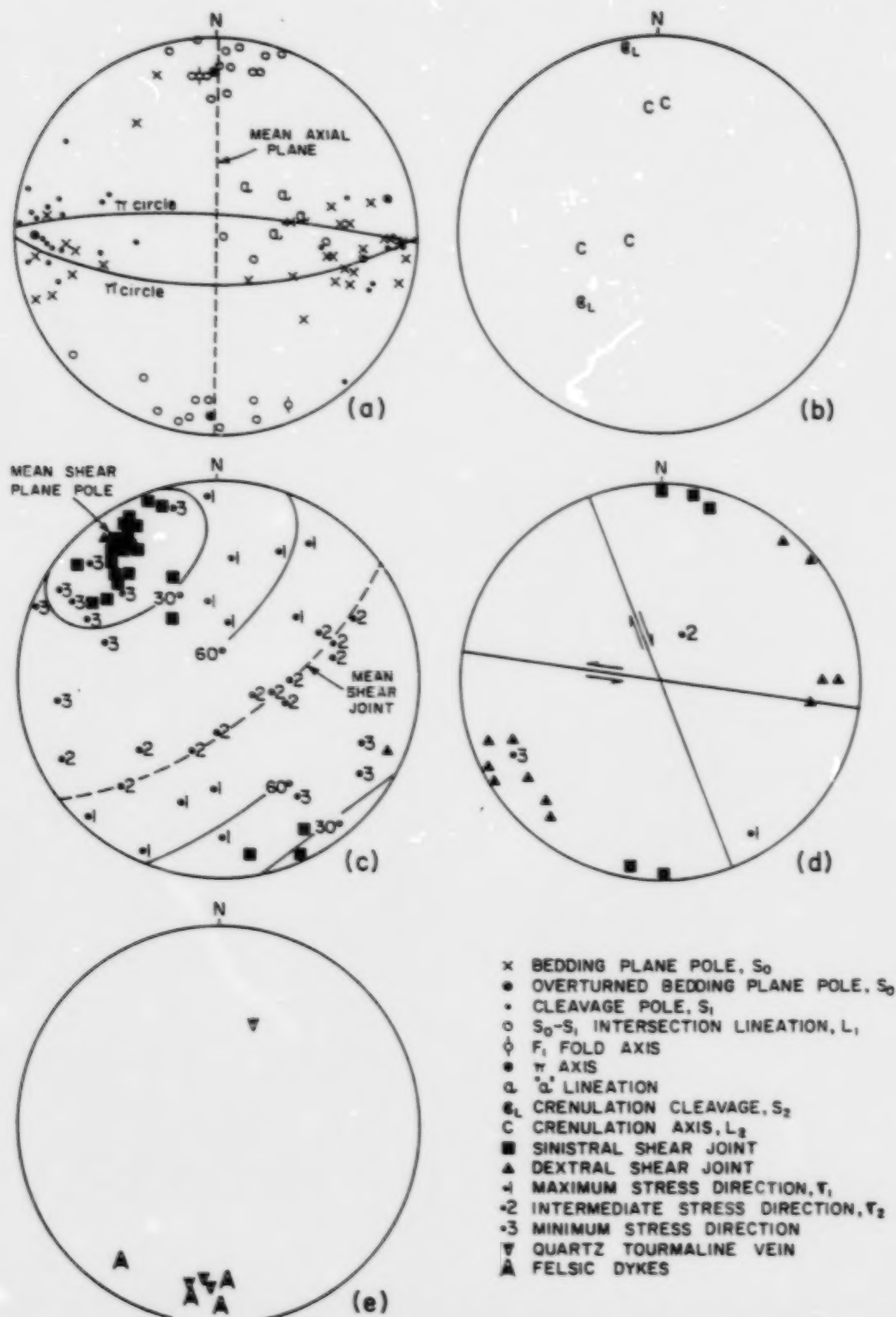


Figure 11-2. Orientations of structures in the Wedgeport area. Lower-hemisphere equal-area projections. a. poles to cleavage; b. crenulation cleavage; c. northeast trending joints; d. northeast-southwest joints; e. dykes and veins (for explanation see text).

continues to the rim or ends at the rim boundary. The rim plagioclase is commonly zoned and the anorthite content decreases against the alkali feldspar;

(c) the rim material in some cases contains inclusions (relicts) of alkali feldspar with the same orientation as the host alkali feldspar.

From the foregoing textural description, it may be concluded that in the unaltered biotite monzogranite, the myrmekite plagioclase between plagioclase and alkali feldspar has nucleated on pre-existing plagioclase and grown into alkali feldspar. Between two adjacent alkali feldspar grains, the sodic plagioclase has nucleated on one of the feldspar grains, adopting its crystallographic orientation and grown into another. Thus, the textural and mineralogical relations associated with the myrmekite in the unaltered biotite monzogranite are consistent with both the exsolution and replacement hypotheses (Phillips, 1974).

'PARAINTRUSIVE' AND MINERALIZED GRANITIC ROCKS

During the geological mapping, it became clear that rocks exposed along the coastline at Pinkney Point are not homogeneous but consist of several varieties, which differ from each other in texture of mineral composition or both. The following varieties have been identified: (a) equigranular granite; (b) porphyritic biotite - muscovite granite; (c) albitized granite; (d) greisenized granite; and (e) greisens. We emphasize that all these rock types are present only on a local scale and the coastal section at Pinkney Point (360 m) is ideally suited to examine the above suite of granitic rocks where transition from one rock type to another occurs within tens of metres. We emphasize that these rocks are genetically related and are products of secondary alterations caused by internally generated fluids.

Equigranular Granite

These granitic rocks are medium grained, xenomorphic to hypidiomorphic, and equigranular but they do contain a few megacrysts of feldspar and quartz. The quartz is light grey and the oxidation of disseminated sulphides imparts a buff or cream colour to the thin rock. In thin section, the feldspars are partly clouded by clay minerals and contain scattered grains of muscovite. K-feldspar is faintly microperthitic, has undulatory extinction, and the grain boundaries range from orthogonal to very irregular and interstitial. Microprobe data (Table 11-1) has shown that this rock type contains four feldspars: (1) larger well zoned sodic oligoclase; (2) smaller unzoned albite, (3) larger sodic K-feldspar; and (4) smaller soda-free K-feldspar. Biotite is anhedral to euhedral and pleochroic in shades of brown. Allanite, zircon, pyrite, pyrrhotite, chalcopyrite, fluorite, cassiterite and scheelite are accessory minerals. A characteristic feature of this rock type is the presence of small (often microscopic) microlitic cavities which are usually filled with quartz, topaz, sericite, chlorite, muscovite and wolframite, and in some cases biotite is also present.

Porphyritic Biotite-Muscovite Granite

In this rock type, large (1-2 cm) crystals of K-feldspar and albite are contained in fine-to medium-grained groundmass (av. 0.1 to 2 mm). The amount and the size of the megacryst vary and, in places, the granite changes to nearly equigranular fine grained rock. Often the marginal parts of the megacrysts of perthitic alkali feldspar contain zonally arranged inclusions of quartz, biotite and muscovite. Also, in the central parts of the megacrysts, discrete, differently oriented plagioclase laths often appear as inclusions. Occasionally, there are micrographic quartz-microcline intergrowths of the radiating fringe type around the euhedral perthitic microcline core. Outwards from the inclusion-rich zone the alkali feldspar has, in many cases, grown anhedrally against the minerals of the groundmass. In the groundmass, turbid plagioclase and quartz occur as the most euhedral minerals. The alkali feldspar may show euhedral outlines against mica, chlorite and topaz. Albitic plagioclase occurs partly as euhedral or subhedral turbid grains, and partly as clear albite rim or grain rows or single grains located usually between turbid plagioclase and K-feldspar or between two adjacent K-feldspar grains. The anorthite content of the plagioclase feldspar is less than five mole percent (Table 11-1). Phyllosilicate minerals occur as small flakes or vein-like bodies replacing feldspar.

Albitized Granite

The albitized granite outcrops at Pinkney Point and is in sharp contact against the Goldenville Formation. The albitized granite is a reddish grey, porphyritic rock which contains angular or corroded megacrysts of albite and quartz in a fine grained groundmass (av. size 0.3-0.5 mm). This rock type is devoid of K-feldspar. The groundmass contains the same minerals and, in addition, dark reddish brown biotite which has been partly altered to green biotite and chlorite. As accessories, the groundmass contains zircon, silverite, anatase, apatite, calcite, muscovite, chlorite, cassiterite and scheelite. Secondary fluorite occurs as thin veins and patches in albite. The albite megacrysts commonly contain drop-like inclusions of quartz and, at the margins, often form micrographic intergrowths of the radiating fringe type. In many cases, the quartz units continue from one grain to another without any change in crystallographic orientations.

Greisenized Granite and Greisens

Greisenized granite occurs usually around the greisen bodies and is characterized by marked hydrothermal alteration dominated by muscovitization and chloritization. Biotite typically forms euhedral to subhedral books scattered evenly through the greisenized granite. The biotite is generally accompanied by a group of minerals which are mainly interstitial or occur with interstitial minerals, namely fluorite, tourmaline, apatite, muscovite, topaz, cassiterite and chlorite, and it is devoid of inclusions of zircon and apatite. Muscovite and chlorite occur as thin rims on biotite, in zones cutting

A

SiO ₂	Al ₂ O ₃	MgO	K ₂ O	CaO	Si	Al	Ca	Na	K	Σ	Or	Ab	An
62.70	22.85	9.50	0.12	4.55	11.149	4.708	0.867	3.275	0.027	15.937	0.653	76.556	20.792
63.72	21.95	9.98	0.13	3.41	11.349	4.607	0.651	3.446	0.030	15.956	0.716	83.515	15.769
64.22	22.63	9.83	0.17	3.94	11.272	4.681	0.741	3.345	0.038	15.953	0.923	81.111	17.876
64.89	22.39	9.89	0.18	3.54	11.355	4.617	0.664	3.355	0.040	15.972	0.990	82.660	16.350
64.19	22.17	9.49	0.17	3.42	11.377	4.630	0.649	3.261	0.038	16.007	0.973	82.581	16.446
62.45	23.60	8.70	0.23	4.99	11.068	4.929	0.948	2.989	0.052	15.997	1.303	74.493	23.754
63.05	23.03	8.83	0.24	4.42	11.192	4.817	0.841	3.039	- .54	16.009	1.301	77.25	21.369
64.15	18.44	1.72	14.40	-	11.951	4.048	-	0.621	3.422	16.000	84.634	15.366	-
64.16	18.40	1.84	14.25	-	11.955	4.040	-	0.665	3.387	15.996	83.593	16.407	-

B

SiO ₂	Al ₂ O ₃	MgO	K ₂ O	CaO	Si	Al	Ca	Na	K	Σ	Or	Ab	An
67.59	19.92	11.48	0.13	0.70	11.860	4.119	0.132	3.906	0.029	15.979	0.716	96.048	3.236
67.46	19.63	11.61	0.13	0.49	11.894	4.078	0.093	3.969	0.029	15.973	0.715	97.022	2.263
68.75	19.85	11.31	0.14	0.44	11.949	4.065	0.082	3.811	0.031	16.014	0.791	97.121	2.088
69.00	19.98	11.65	0.20	0.21	11.937	4.073	0.039	3.908	0.044	16.010	1.106	97.919	0.975
68.75	20.19	11.71	0.06	0.66	11.873	4.109	0.122	3.921	0.013	15.982	0.326	96.663	3.011
68.78	20.03	11.65	0.11	0.54	11.902	4.084	0.100	3.909	0.024	15.987	0.602	96.916	2.482

C

SiO ₂	Al ₂ O ₃	MgO	K ₂ O	CaO	Si	Al	Ca	Na	K	Σ	Or	Ab	An
68.48	19.52	11.72	0.07	0.11	11.975	4.022	0.021	3.973	0.016	15.997	0.389	99.097	0.514
64.37	22.29	9.94	0.06	3.41	11.350	4.631	0.644	3.398	0.013	15.982	0.333	83.784	15.884
65.14	21.61	10.30	0.12	2.94	11.472	4.485	0.555	3.517	0.027	15.957	0.658	85.807	13.535
68.06	19.57	11.65	0.07	0.19	11.949	4.049	0.036	3.966	0.016	15.998	0.390	98.720	0.890
67.66	19.51	11.75	0.10	0.27	11.925	4.052	0.051	4.015	0.022	15.977	0.009	98.203	1.247
68.49	18.25	-	16.30	-	12.022	4.009	-	-	3.876	16.032	100.00	-	-
65.69	18.42	1.92	13.89	-	12.034	3.976	-	0.682	3.246	16.011	82.637	17.363	0.00
64.53	18.44	1.31	14.98	0.13	11.961	4.028	0.026	0.471	3.542	15.989	87.702	11.658	0.639

D

SiO ₂	Al ₂ O ₃	MgO	K ₂ O	CaO	Si	Al	Ca	Na	K	Σ	Or	Ab	An
68.57	19.60	11.81	0.07	-	11.971	4.032	-	3.938	0.016	16.004	0.388	99.612	0.000
68.84	19.85	11.57	0.11	0.05	11.959	4.063	0.009	3.897	0.024	16.038	0.620	99.143	0.237
69.49	19.53	11.57	0.07	0.07	12.024	3.982	0.013	3.982	0.015	16.021	0.395	99.273	0.332
68.57	19.98	12.00	0.06	0.02	11.915	4.091	0.004	4.043	0.013	16.006	0.328	99.581	0.092

Table 11-1. Electron microprobe analyses of feldspars from the granitoid rocks of the Wedgeport pluton. Number of ions in the structure is calculated on the basis of 32 oxygen.

biotite crystals and also as flakes intergrown with biotite. Accessory minerals consist of tourmaline, sphalerite, chalcopryrite, arsenopyrite, cassiterite, wolframite, scheelite, native bismuth, argentite, pyrite and pyrrhotite.

The greisen zones range in width from a few centimetres to over three metres, and compositionally are extremely varied. Some are dominated by muscovite while others are essentially quartz veins. The sulphide content of the greisens range from less than five percent to more than 70 percent. The major gangue minerals of the greisens are quartz and phyllosilicates together with topaz, fluorite, beryl, tourmaline and carbonates. The common ore minerals are the same as those observed in the greisenized granite.

GEOCHEMISTRY OF THE GRANITOID ROCKS

Thirty-seven samples, representing unaltered rocks, altered rocks and mineralized rocks from the Wedgeport Pluton were analyzed for major and trace elements. It was necessary to log transform the data. The correlation matrix (Fig. 11-3) shows the significant (99.9% level of significance) covariance of major and trace elements. Important relationships are that both uranium and thorium correlate positively with SiO_2 and most other elements correlate negatively with SiO_2 (e.g. TiO_2 , Al_2O_3 , FeO , MgO , CaO , P_2O_5 , Ba , Li , Sn and Zn). Also, uranium and thorium correlate with each other positively. In addition, to the gross chemical features noted above, we have plotted various binary diagrams (Figs. 11-4 to 11-8) using the differentiation index (D.I.) simply because this is a good summary index of bulk composition. The essential features of these diagrams are briefly discussed in the following section.

As to be expected, SiO_2 increase with DI in the three groups of rocks and the trend lines are almost parallel to each other. Similarly, TiO_2 decreases with increasing DI. However, the mineralized rocks are characterized by low TiO_2 and they define a trend which is markedly different than those defined by the unaltered or altered rocks. The variation diagrams involving other elements such as MgO , FeO , CaO , and P_2O_5 define similar characteristics. The plot involving normative colour index (CI) and DI clearly shows the unaltered rocks have a higher index whereas the altered rocks have distinctly lower indexes, and, in both cases, the CI decreases with increasing DI. The mineralized rocks have a CI of 0.2 to 5.5 and this is again vastly different from the unaltered and altered rocks. The plot involving alumina ratio (normative corundum) and DI shows the rocks are generally characterized by a range of alumina ratio higher than 1.10, attesting to the peraluminous nature, but decreasing with increasing DI for the unaltered rocks and increasing with DI for altered rocks. Also, for equivalent alumina ratio the unaltered rocks have less normative corundum than the altered rocks. The alumina ratio and normative corundum values decrease with increasing DI for the mineralized rocks and essentially it reflects whether it is a mica greisen or a quartz greisen. The alkalis

vs. DI plot defines a pattern which reflects their mobility during hydrothermal alteration, and the role of albitization and formation of secondary albite during mineralization. It is clear that on an average, Na_2O remains more or less constant in the unaltered rocks but shows a positive correlation with DI for the altered and mineralized rocks.

The binary diagram of the total volatile contents of the rock against the DI shows clearly in the unaltered rocks it is relatively constant but for the altered rock it increases progressively and is characterized by a strong positive correlation. For the mineralized rocks, the mica greisens are higher in volatiles where as the quartz greisens are lower, thus defining a negative correlation with DI. The Cu/Zn ratio decreases with increasing DI for the unaltered rocks whereas for the hydrothermally altered rock it increases. Thus it is negatively correlated with DI for the unaltered rocks and the reverse for the altered rocks. The mineralized rocks define a U-shaped distribution where it clearly illustrates the dominance of copper over zinc and this is consistent with the mineralogical observation that in the greisen chalcopryrite is prevalent. The distribution of ratioelements as per airborne gamma ray spectrometry is shown in Figure 11-9a. It clearly shows that the Wedgeport pluton is characterized by high uranium and thorium counts but the eU/eTh ratio is low and in fact this ratio is not very different than the surrounding Cambro-Ordovician metasediments. Figure 11-9b shows the distribution of uranium and thorium for the unaltered, altered and mineralized rocks. The unaltered rocks are characterized by comparatively low uranium and high thorium values, and shows a positive correlation. The altered rocks show a much wider range in uranium values and though it is positively correlated with uranium, the slope is different from the unaltered rocks, perhaps consistent with the concept of mobilization of uranium during hydrothermal alteration. The uranium values range from 7.2 to 22 ppm for the mineralized rocks but range for thorium is less than 6 ppm. In summary, the variation diagrams support a general evolutionary trend from magmatic to hydrothermal to metallization, and perhaps an ideal illustration of the various processes discussed by Clarke and Chatterjee (1985) at magmatic and post magmatic stages related to the intrusion of granite bodies.

CONCLUSIONS

The Carboniferous Wedgeport Pluton shows sharp crosscutting contacts, hornfels aureoles, microlitic cavities, aplite-pegmatite segregations and porphyritic rocks with aplitic groundmass, and these features are consistent with an epizonal emplacement environment.

Continuous variations in the ferro-magnesian elements (e.g. TiO_2 , MgO) support the fractionation model for the evolution of the granitic rocks. It also explains the uranium element variation where it could be argued that it was preferentially incorporated in early crystallizing phases (e.g. Th in biotite). Though fractionation explains many geochemical features,

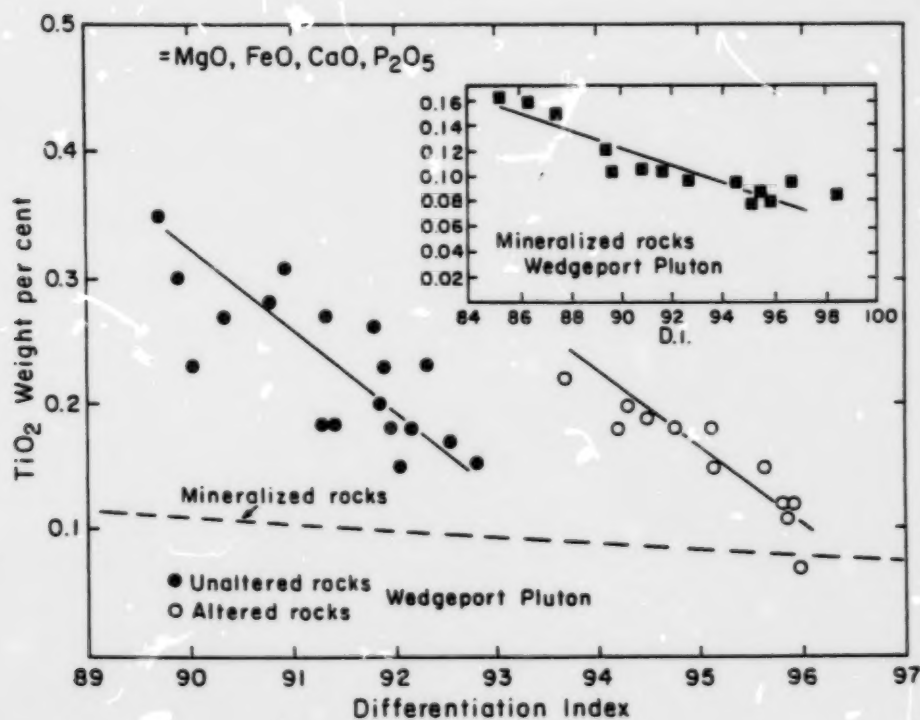
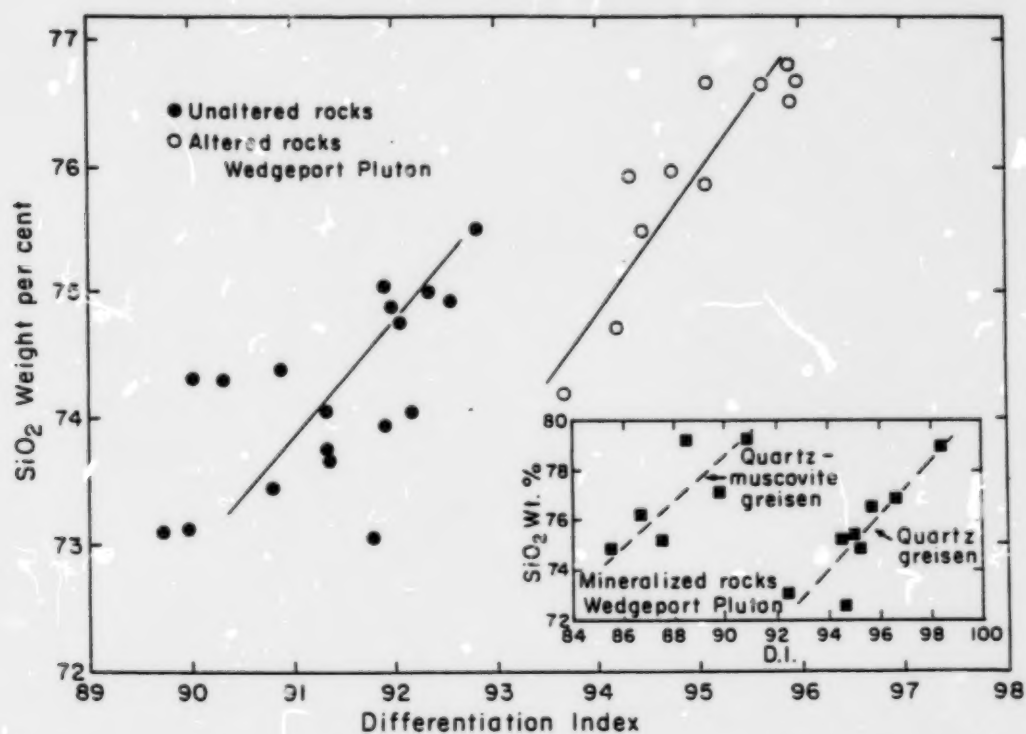


Figure 11-4. Differentiation index vs SiO₂ and TiO₂ variation diagram for the granitoid rocks of the Wedgeport pluton.

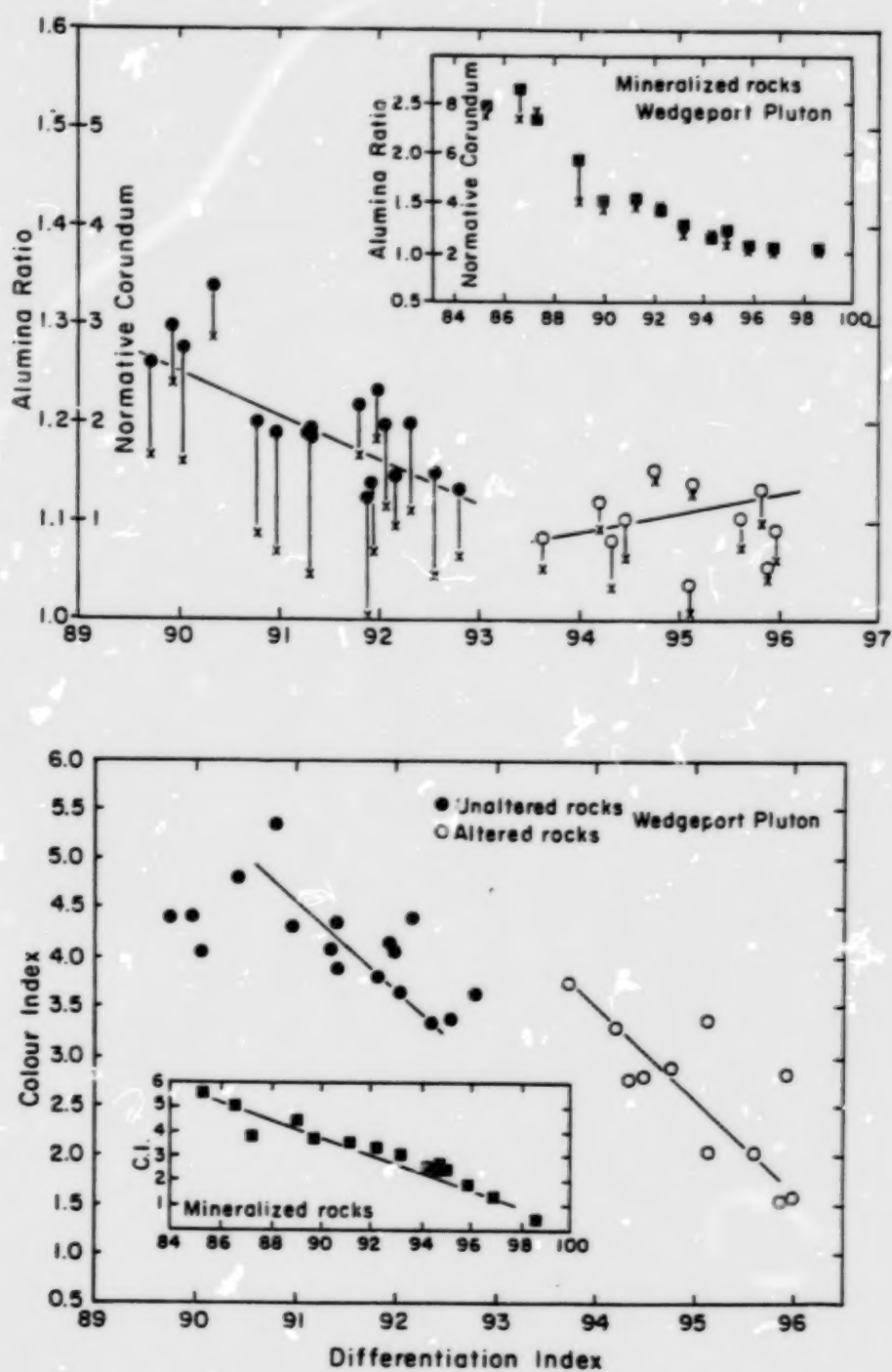


Figure 11-5. Differentiation index vs alumina ratio and normative Colour Index variation diagram for the granitoid rocks of the Wedgeport pluton.

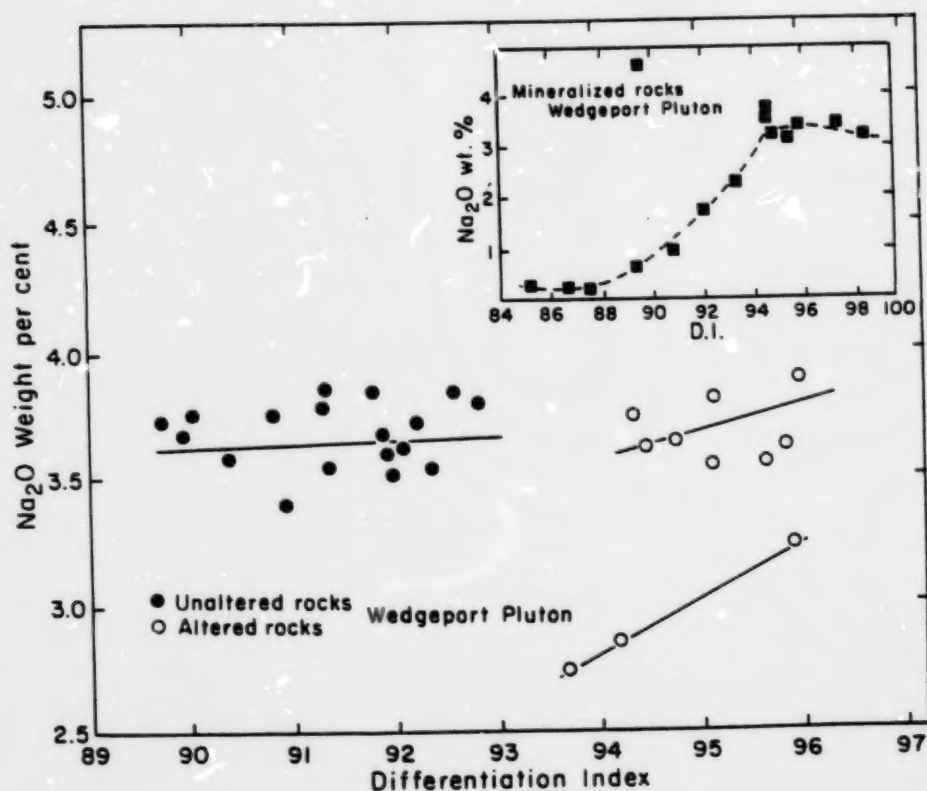
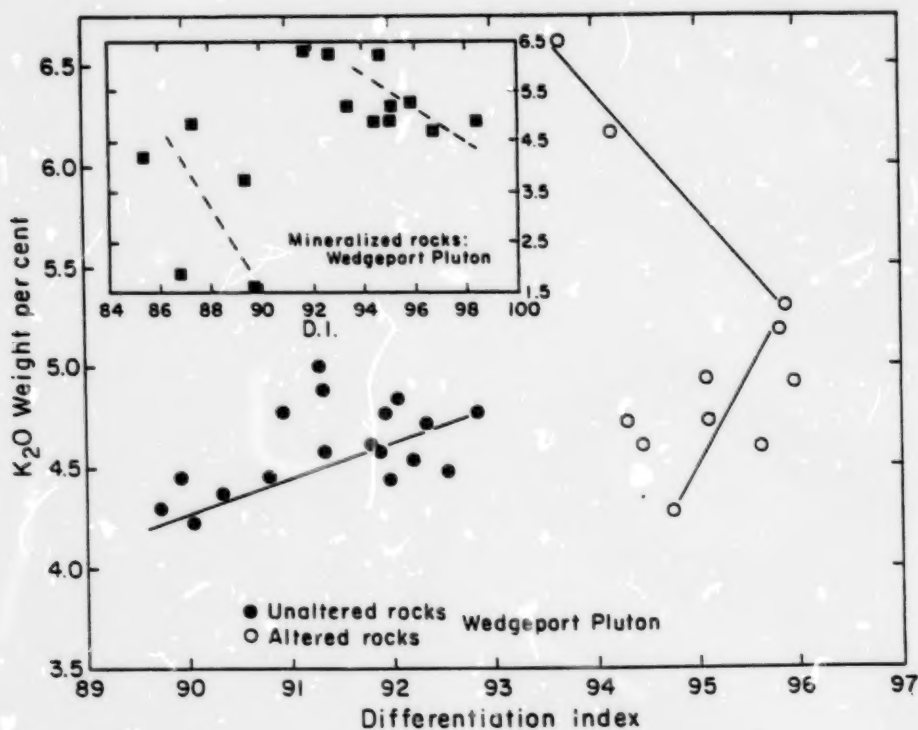


Figure 11-6. Differentiation index vs alkalis variation diagram for the granitoid rocks of the Wedgeport pluton.

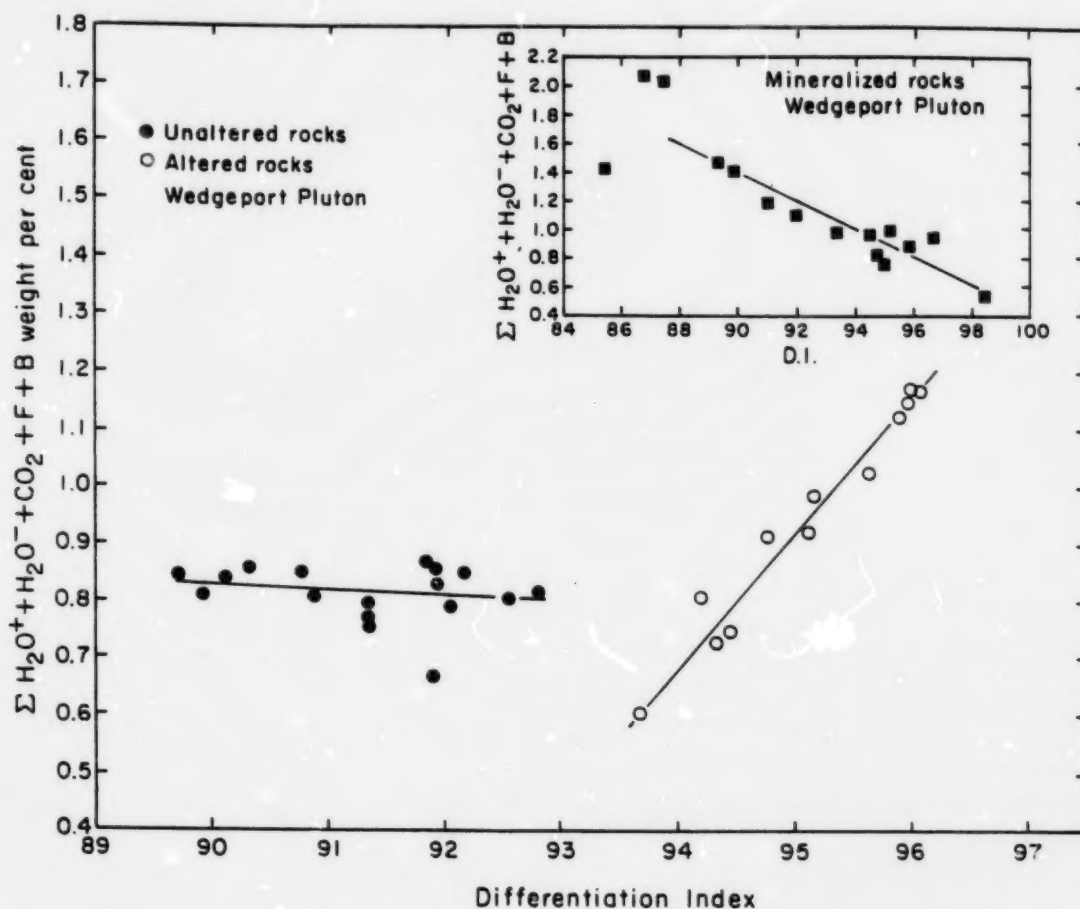


Figure 11-7. Differentiation index vs Σ volatiles variation diagram for the granitoid rocks of the Wedgeport pluton.

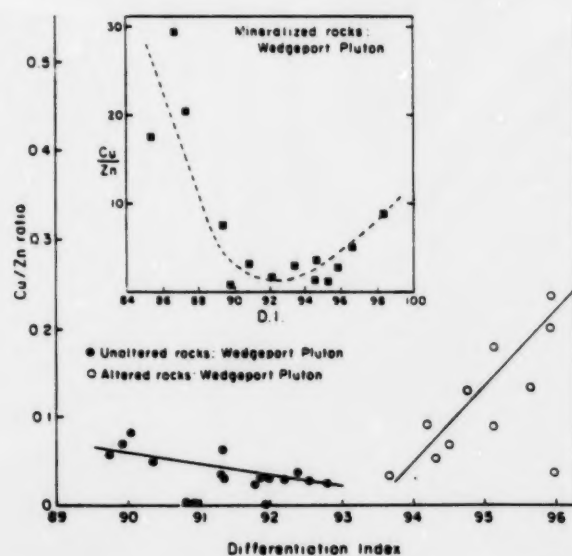


Figure 11-8. Differentiation index vs Cu/Zn ratio variation diagram for the granitoid rocks of the Wedgeport pluton.

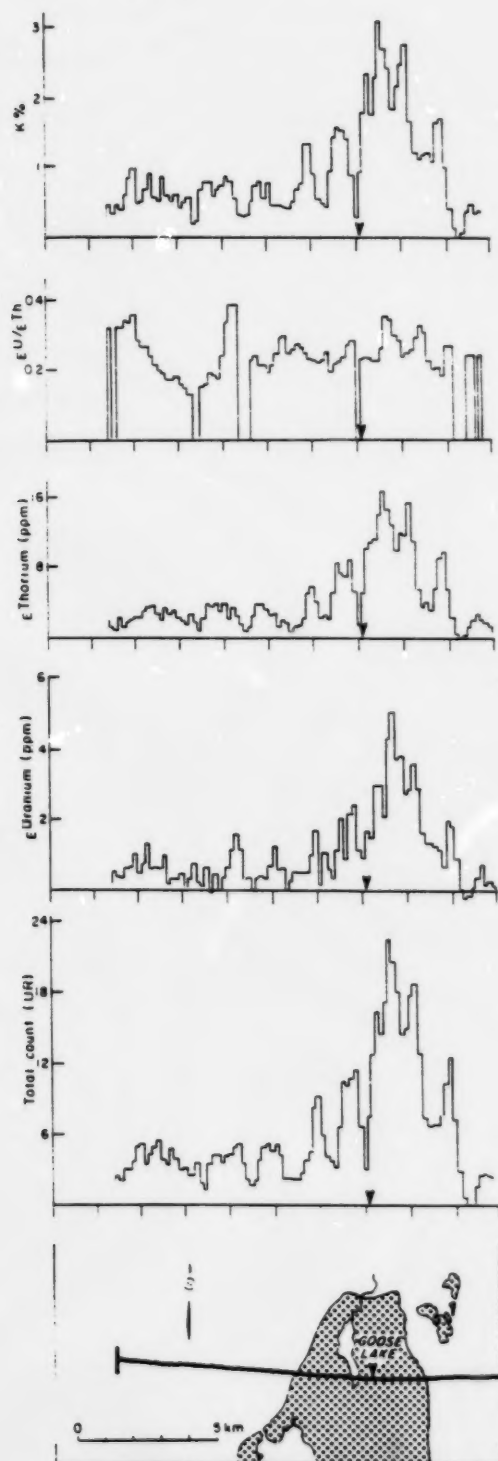


Figure 11-9. (a) Airborne gamma ray response across the Wedgeport pluton.

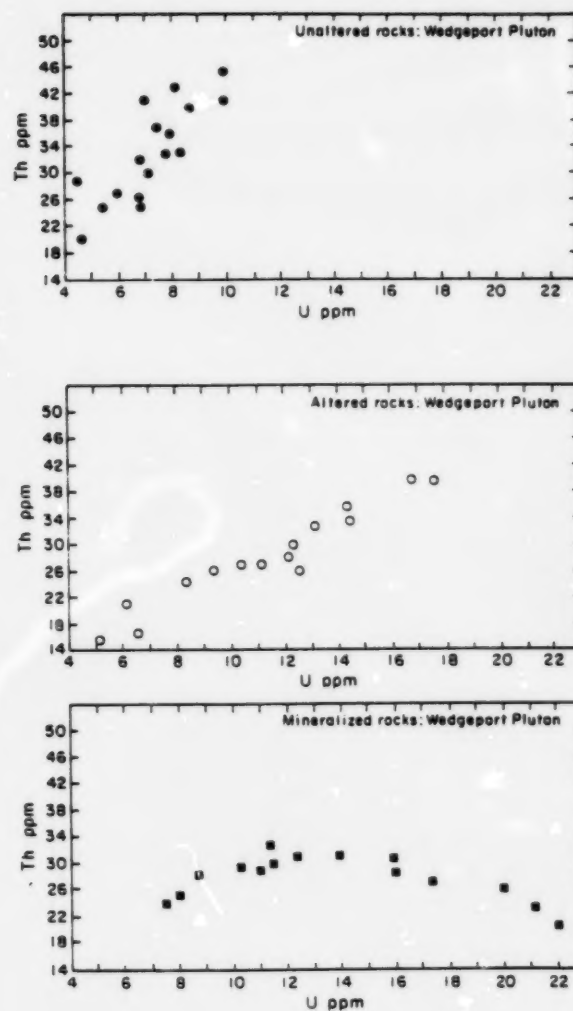


Figure 11-9. (b) U-TH variation diagram for the unaltered rocks, altered and mineralized rocks of the Wedgeport pluton.

it does not readily explain either the degree or ratios of heavy metal (e.g. Cu/Zn) variations. Widespread alteration, miarolitic cavities and the presence of volatile rich greisens all suggest that a coexisting volatile phase was important in the evolution of the rocks of the Wedgeport Pluton. The mineralogical and geochemical data are consistent with the classical pneumatolytic hydrothermal origin for greisen related polymetallic mineralization, that appears to be final product of fractional crystallization of granitic magmas, evolution of a volatile-rich system and timely fracturing of pluton margins to allow escape of a co-existing volatile phase.

CHAPTER 12

ALKALI METASOMATISM ASSOCIATED WITH W-Sn-Ag-Au MINERALIZATION
IN THE GRANITOID ROCKS OF THE WESTFIELD-CALEDONIA AREA,
QUEENS COUNTY, NOVA SCOTIA

G. A. O'REILLY

O'Reilly, G. A., Alkali metasomatism associated with W-Sn-Ag-Au mineralization in the granitoid rocks of the Westfield-Caledonia area, Queens County, Nova Scotia; in Guide to the Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

ABSTRACT

The Westfield-Caledonia area of Queens County, Nova Scotia, consists of Cambro-Ordovician metasedimentary rocks of the Meguma Group intruded by the peraluminous South Mountain Batholith (SMB). The granitoid rocks consist of a coarse grained porphyritic granodiorite-monzogranite with a paraintrusive suite of rocks developed both within a peripheral offshoot of the Batholith and a series of composite granitoid dykes intruded into the exocontact zone. In the endocontact zone the paraintrusive suite consists of altered granodiorite, K-feldspathized albite-bearing porphyry, albitized porphyry, greisenized granitoids and quartz-muscovite greisen. In the exocontact zone a dyke of quartz-albite-chlorite-pyrite porphyry and a composite granitoid dyke are found. The composite dyke consists of a quartz-albite rhyolitic phase, coarse grained albitite phase and fine grained albitized breccia phase. This paraintrusive suite of granitoid rocks has developed by interaction of fluid phase with both the previously crystallized granodiorite of the Batholith and residual magma.

Within the Westfield-Caledonia area, W-Sn-Mo-Ag mineralization occurs in the endocontact paraintrusive phases as disseminations in highly greisenized zones and pods. Within the exocontact zone W-Sn-Cu-Pb-Zn-Ag-Au occurs predominantly within a series of quartz fissure veins and their associated greisen haloes related to the paraintrusive rocks. One such vein, the Jumbo Vein, is an 8 to 22 m thick quartz vein and silicified country rock zone formed along a northwest trending shear. This zone is known to host Au-Ag mineralization associated with polymetallic sulphides.

Major element geochemistry indicates both potassic and sodic alteration occurred within the paraintrusive phases but with an apparent dominance of soda metasomatism. Trace element geochemistry has defined a granophile element specialization assemblage of Sn, W, B, F, Cu, Zn, Au and Ag. An interesting variation of this element assemblage from that of other analyzed Meguma Zone paraintrusive suites is a lack of Rb-U specialization. The absence of Rb and U specialization and the observed field relations of the Westfield suite suggest these rocks developed after early separation of a fluid phase from a granodioritic magma. This resulting more primitive fluid phase was less enriched in Rb and U, and interacted with crystallized granodiorite and residual magma to produce the observed suite. To facilitate early separation of this fluid phase the magma may have been enriched in CO₂ and Cl. Petrography verified CO₂ was dominant within this system and a Au specialization within these rocks suggests a high Cl activity may also have existed.

INTRODUCTION

The Westfield-Caledonia area is located in the northeast of Queens County, south-central Nova Scotia (Fig. 1). At Westfield, W-Sn-Ag-Au mineralization is localized in granitoid rocks and associated alteration zones and quartz veins

exposed along the contact zone of the SMB with Meguma Group metasedimentary rocks. A short distance to the southwest Sn-W mineralization occurs in quartz veins and composite granitoid dykes intruded into Meguma Group metasedimentary rocks.

Gold and silver has been known to occur at Westfield since 1888 but this particular deposit never achieved more than prospect status. The precious metal mineralization is hosted by what Malcolm (1976) termed large quartz saddle reefs formed in Upper Halifax Formation slates. The models for formation of gold deposits in Nova Scotia in the late 1800-early 1900 era led explorationists and prospectors to classify the veins in this area with the many saddle reef gold deposits found throughout mainland Nova Scotia. This bias held even though the veins in this area display obvious geological, structural and mineralogical differences from most of the gold mining districts. There was no recognized correlation of this particular deposit genetically or spatially with the granitoid rocks which outcrop a short distance to the north.

Shell Canada Resources, during regional surveys, discovered Sn-W mineralized boulders and outcrops in the Caledonia area in 1977 (Wilson 1978). During the following years, they carried out follow-up geochemistry and preliminary diamond drilling and concluded the occurrences are likely associated with a buried granitic source. Billiton Canada Limited optioned the ground in 1982 and continued detailed surveys on the property. The Caledonia area is at present held by Billiton Canada Limited although no exploration is being carried out.

REGIONAL GEOLOGY OF THE MEGUMA ZONE

Mainland Nova Scotia south of the Cobequid-Chedabucto Fault Zone is underlain by a sequence of Cambrian-Early Devonian metasedimentary - metavolcanic rocks and a suite of Devonian-Carboniferous peraluminous granitoid rocks, unconformably overlain by sedimentary rocks of Carboniferous and Triassic age. This region of Nova Scotia is known as the Meguma Zone (Williams, 1978).

Much of the metasedimentary pile comprising the Meguma Zone is a thick turbidite sequence known as the Meguma Group. The Goldenville Formation and conformably overlying Halifax Formation make up the Meguma Group. The Goldenville Formation consists predominantly of metawackes and quartzite with minor slate and schist while the Halifax Formation consists predominantly of slate and schist with minor quartzite (Taylor, 1969). Toward the northwest, conformably overlying the deep water deposited Meguma Group, are the shallow water facies rocks of the White Rock, Kentville, New Canaan and Torbrook Formations. The metasedimentary rocks range in age from Cambro-Ordovician (Meguma Group) to Early Devonian (Torbrook Formation). The entire sequence was deformed into upright northeast trending folds with the onset of the Acadian Orogeny (Fyson, 1966). Greenschist-amphibolite regional metamorphism accompanied the middle-late stages of this orogeny (Keppie and Muecke, 1979).

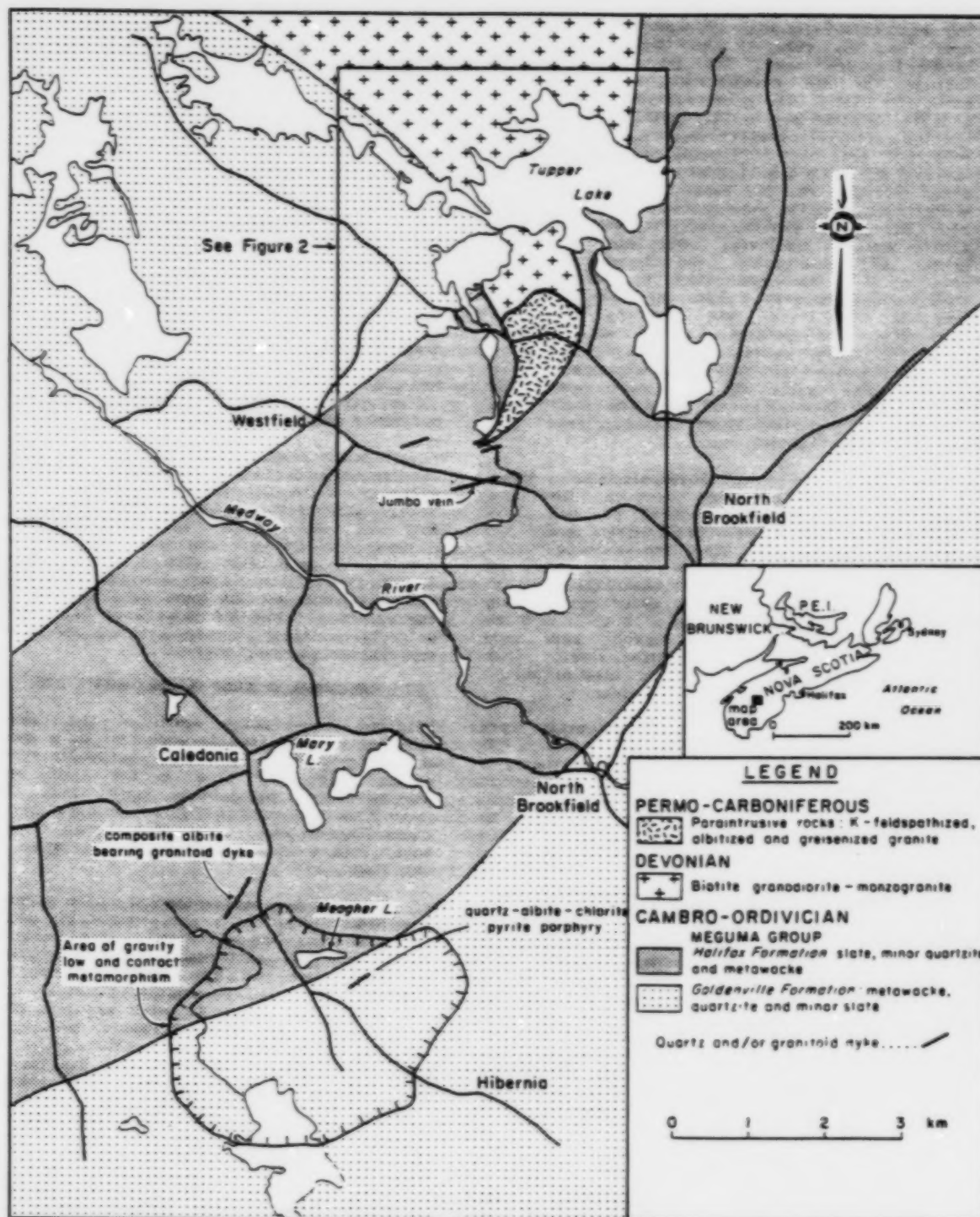


Figure 12-1. Geology of the Westfield-Caledonia area of Queens County, Nova Scotia.

A considerable portion of the Meguma Zone is underlain by a cogenetic suite of peraluminous granitoid rocks. The largest of these massifs outcrops over 10 000 km² and has been named the SMB (McKenzie and Clarke, 1975). A number of similar but smaller granitoid plutons occur throughout the rest of the Meguma Zone, some of which have been examined by various workers (Smith, 1974; Albuquerque 1977; MacDonald and Clarke, in press). The granitic complexes represent a cogenetic suite ranging from tonalite to leucogranite with granodiorite and monzogranite comprising 85-90 per cent of present day exposure.

Albuquerque (1977) invoked partial melting of the Meguma Group as the source of the parent magma for granitoids in the southern Meguma Zone. Clarke and Halliday (1980) used strontium isotope data to show the Meguma Group as exposed at surface could not be the sole source of the parent magma of the SMB. They do concede though that the source region must have involved a large volume of metasediment.

The age of the granite rocks is defined on field evidence to be younger than the Early Devonian Torbrook Formation as granodiorite of the SMB has posttectonically intruded this formation. Radiometric dating of the granitoids has placed the bulk of intrusion of the cogenetic suite in the 355-375 Ma interval (Cormier and Smith, 1973; Clarke and Halliday, 1980; Reynolds et al., 1981). Chatterjee and Muecke (1982) have documented the presence of a paraintrusive suite of rocks within the cogenetic suite of the SMB. This paraintrusive suite is the result of interaction of a fluid phase with either residual magma or previously crystallized rock or both. This fluid phase was enriched (specialized) in such elements as Sn, U, Rb, Li, F, Be and Cs, and has resulted in formation of biotite leucogranite, sericitized granite, albitite and various greisens. Although volumetrically these "specialized" paraintrusive rocks only represent a small portion of the granitoid complexes, they are of special interest due to their close spatial and genetic association with significant Sn-U mineralizations in the Meguma Zone (Chatterjee and Muecke, 1982; Richardson et al., 1982; O'Reilly et al. 1985).

Radiometric age determinations of samples from the paraintrusive suite commonly yield apparent ages in the 260-340 Ma Hercynian interval (Reynolds et al. 1981; Chatterjee, 1983; O'Reilly et al., 1985). These younger dates are anomalous relative to the Middle Devonian dates (360-375 Ma) commonly obtained from phases of the cogenetic suite. This time discrepancy is believed to be the result of a Permo-Carboniferous event causing either a separate magmatism or thermotectonism resulting in overprinting of precystallized granite rocks. The nature of this time discrepancy is at present not fully understood but is important considering its economic significance.

GEOLOGY OF THE WESTFIELD-CALEDONIA AREA

The Westfield-Caledonia area is underlain by metasedimentary rocks of the Meguma Group in the

south and by granite rocks of the SMB in the north (Figs. 12-1, 12-2). The areas of study in this paper are a peripheral protrusion of the Batholith and its exocontact zone immediately east of Westfield, and an area of metasedimentary Meguma Group rocks hosting dykes of highly evolved paraintrusive rocks south of the village of Caledonia.

METASEDIMENTARY ROCKS

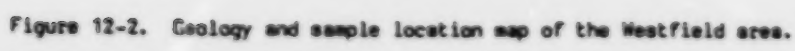
Within the study area the Meguma Group consists of both the Goldenville and Halifax Formations. The Goldenville Formation consists predominantly of fine grained massive beds of quartzite, quartz wacke and greywacke. Locally grey-green slate horizons occur. In the Meagher Lake - Mary Lake area near Caledonia many small carbonate concretions and calc-silicate horizons occur. The abundance of these calc-silicate horizons peaks near the Goldenville-Halifax Formation contact. This is a common feature throughout the Meguma Zone (P. K. Smith, personal communication).

The Halifax Formation in this area conformably overlies the Goldenville Formation with the contact represented by a thick transitional zone of predominantly green siliceous slate grading upward into grey-green slate and finally the typical grey-dark grey sulphide-bearing slates of the Halifax Formation. A similar but thicker transition sequence is described by O'Brien (1985) in the Bridgewater area to the southeast. In this paper the transitional unit is included as Halifax Formation on Figure 12-1.

The Meguma Group in this area has been deformed into tight northeast trending folds. South and west of Upper Lake shear related zones of dislocation parallel the regional fold axes. At Westfield at least two of these zones have been infilled by quartz and are silicified suggesting this generation of shear may have been important in localization of mineralization. The metasedimentary rocks of this region of the Meguma Zone have been raised to chlorite grade of regional metamorphism. Two areas of higher grade metamorphism have been found. Along the Batholith - metasediment contact the Meguma Group country rocks have undergone posttectonic contact metamorphism associated with the magmatic event. In the area of Meagher Lake, south of Caledonia, similar contact hornfels occur but as the Batholith is some kilometres away this area of metamorphism is anomalous. The coincidence of this area of anomalous metamorphism with occurrences of dykes of highly evolved paraintrusive rock and a bullseye shaped gravity low lead Shell Canada Resources to conclude this area is underlain at shallow depth by a granitic cupola (Wilson, 1978; MacGillivray, 1982, 1983; Fig. 12-1).

GRANITOID ROCKS

The cogenetic region of the SMB exposed in this area consists predominantly of one phase, a porphyritic biotite granodiorite monzogranite. In hand specimen this rock is typical of this phase throughout the Batholith, containing subhedral to euhedral phenocrysts of orthoclase in a coarse grained hypidiomorphic granular matrix of quartz, plagioclase (An₂₀₋₂₅), lesser



orthoclase, biotite and muscovite. Accessories include zircon, apatite, cordierite and tourmaline.

Paraintrusive Rocks

Mineralization in the Westfield-Caledonia area is associated spatially and genetically with a variable complex of paraintrusive rocks. At Westfield the paraintrusive suite occurs at the periphery of a lobe of the Batholith extending south of Ivor Lake (Fig. 12-2). In the area north and south of the Rosette Road, the porphyritic granodiorite apparently grades texturally into altered granodiorite and well developed K-feldspathized porphyry. The altered granodiorite in hand specimen resembles the granodiorite except for chloritization of biotite and sericitization of feldspar. The K-feldspathized porphyry is grey-cream in colour with phenocrysts of albite, quartz, secondary biotite or chlorite (or both) set in a matrix of interstitial orthoclase, quartz, albite, sericite, calcite and pyrite.

Toward the periphery of the body the K-feldspathized porphyry grades into albitized porphyry. This albitized phase texturally resembles the K-feldspathized phase differing only in absence of potassic feldspar. Greisenized granite is characteristically found at the outermost extremities of the porphyry body and locally along fractures within the porphyry. Increase of progressive metamorphism has resulted in depletion of albite with increase in modal muscovite in this phase. It is within these rocks that most endocontact W-Sn mineralization occurs.

The paraintrusive rocks of the Caledonia area occur as two dyke-like complexes intruded into cordierite-bearing hornfels of the Meguma Group (Fig. 12-1). South of Meagher Lake a purple-grey quartz-albite-chlorite-pyrite porphyry forms a small (2m) wide dyke. The intrusion is homogeneous except for some chilling toward its margins. Pyrite occurs both as discrete blebs within the groundmass and as crystals mantled by albite.

Northwest of Meagher Lake a complex of various paraintrusive rocks forms a composite dyke approximately 200 metres long in a north-northeast direction. Within the dyke occurs medium to coarse-grained albitite, a quartz-albite rhyolitic porphyry and a tourmaline bearing breccia with abundant angular Meguma Group fragments. The interrelationships of the phases is complex but some features were observed. Coarse grained albitite occurs both as a phase forming sections of the dyke and as rounded clasts in the quartz-albite porphyry. This indicates the quartz-albite porphyry is younger than the coarse grained albitite phase. The existence of rounded coarse grained albitite clasts within the porphyry suggests the former has been brought up from depth perhaps from the underlying cupola. The coarse grained albitite texturally and compositionally resembles the albitized porphyry exposed in the Westfield area. This suggests the albitite dyke phase and rounded clasts may have been derived from a source similar to the albitized porphyry.

The quartz-albite rhyolite porphyry was not observed in contact with the coarse grained albitite phase but conceivably the former is a finer grained chilled variety of the albitite. This is supported by the similar chemistry of these phases described in the lithogeochemistry section to follow. Rounded fragments of the quartz-albite porphyry and angular Meguma Group fragments occur in a fine grained tourmaline-bearing breccia phase in a number of outcrops of the composite dyke. This breccia phase is therefore younger than the quartz-albite rhyolitic porphyry, and obviously explosive and volatile rich in nature. Much of the significant mineralization associated with the dyke complex occurs within this breccia or in quartz vein offshoots from it.

MINERALIZATION

Endocontact

Within the paraintrusive suite of rocks scheelite, wolframite, cassiterite, molybdenite, arsenopyrite and pyrite occur as disseminations and vug fillings in the greisenized granitoids, and in quartz vein stockworks and fissure veins. The mineralization is sporadic although a number of highly greisenized zones are exposed, e.g., along the Westfield River north of the bridge on the North Brookfield Road, and in the woods north of the junction of the Rosette and Harlow Roads (Fig. 12-2).

Exocontact

Within the exocontact zone of the cupola of paraintrusive rocks exposed along the Westfield River are found a number of quartz fissure veins and associated greisen haloes in Halifax Formation slate. Along the river bed are exposed several quartz fissure veins and quartz stockwork containing scheelite, cassiterite, Cu-Pb-Zn sulphides, arsenopyrite, pyrite and marcasite. South of this river zone is found the Jumbo Vein, an 8-22 m wide quartz fissure vein hosted in a greisenized and silicified shear zone. Most past exploration activity in this area was concerned with this fissure vein zone to examine its gold potential. Most interest was directed toward the precious metal potential of the quartz vein proper with little attention paid to its silicified and greisenized halo.

A short distance west of the Jumbo Vein and the river zones, there exists another quartz and silicified zone similar but much smaller (1-2m) than the Jumbo. This zone is exposed along the north edge of the municipal garbage dump site and contains some sporadic disseminations of scheelite and arsenopyrite. This zone lacks the significant greisenized halo of the Jumbo Vein and the river zones.

The Caledonia area is considered an exocontact zone due to the coincidence of a number of features outlined earlier which suggest the existence of a buried granitic cupola. In this area, MacGillivray (1983) described greisen veins and associated breccia pipes containing polymetallic sulphides, scheelite and cassiterite, and bull quartz veins containing

spectacular crystals of cassiterite and scheelite along the vein-wallrock contact. In the Caledonia area most mineralization occurs along or within the composite granitoid dyke complex.

AGE RELATIONS

Clarke and Halliday (1980) have reported a Rb/Sr mineral and whole rock isochron apparent age of 371.8 ± 2.2 Ma for the biotite granodiorite phase of the SMB. This date is assumed to represent the age of this phase in the Westfield area as consistently the granodiorite of the Batholith has been dated in the 370 Ma range. O'Reilly et al. (1985) have reported a Rb/Sr whole rock isochron age of 270 ± 14 Ma for K-feldspathized porphyry samples from the Rosette Road. This phase apparently has a gradational relationship with the biotite granodiorite outcropping to the north therefore the age discrepancy of 100 Ma between these phases is somewhat perplexing. A logical explanation would be that the processes resulting in K-feldspathization and age updating of the granodiorite are the result of a Permo-Carboniferous thermotectonic event in this area. Another explanation is that the observation of a gradational relationship between these phases is wrong. The observed coarsening of grain size within the porphyry toward the granodiorite may be misleading in which case the porphyry could be a separate magmatic intrusion. At present this aspect is unresolved but the author favours a gradational relationship.

LITHOGEOCHEMISTRY

Forty-five samples from the cogenetic and para-intrusive suites from the Westfield-Caledonia area were analysed for major and trace elements (Table 12-1). The samples have been classified according to rock types as described in the previous section. For simplicity and to reduce the number of symbol types required on elemental plots used in this section, some of the para-intrusive phases have been grouped according to the dominant alkali species of alteration. A potassic facies includes rocks of dominantly potassic alteration and a sodic facies for rocks of predominant albitic alteration.

In addition to the major and trace element analyses, twenty samples of various greisens and associated quartz veins were submitted for selected trace element assays (Table 12-2). The assayed samples consist of outcrop, waste rock dump and boulder samples from the Westfield area.

MAJOR ELEMENTS

Examination of the major element data in Table 12-1 show that the porphyritic granodiorite phase of the Westfield area compares in overall chemistry with that of biotite granodiorite for the SMB reported by Smith (1974), and McKenzie and Clarke (1975). The para-intrusive suite of rocks in the Westfield-Caledonia area displays major elemental variations of alkalis indicating potassic and sodic metasomatism predominated. A ternary plot of K_2O-Na_2O-CaO (Fig. 12-3) compares the various cogenetic and para-intrusive phases with the fields of granodiorite and monzogranite for the South Mountain and

Musquodoboit Batholiths of the Meguma Zone. The plot shows the predominance of a strong albitization trend of the sodic facies rocks in the Westfield-Caledonia area. The potassic facies suite shows some K_2O enrichment relative to the cogenetic granodiorite and monzogranite. It should be noted that although the potassic facies rock types are orthoclase- and sericite-bearing they also contain considerable albite (An_{1-3}). The coincidence of both the sodic and potassic species in the alteration fluids, causing the potassic facies rocks, offsets the K_2O to Na_2O proportion thus showing only an overall slight K_2O enrichment on Figure 12-3. The sodic facies rocks are potassic feldspar free, this being reflected in their overall strong Na_2O trend in Figure 12-3. The greisenized granitoid samples show strong K_2O enrichment as these rocks are tending toward a quartz and muscovite assemblage with both albite and orthoclase unstable in the K_2O -rich greisenizing fluids.

TRACE ELEMENTS

Again as with the major elements, the cogenetic porphyritic granodiorite of the Westfield area has trace element levels similar to those for granodiorite of the SMB (McKenzie and Clarke, 1975; Chatterjee and Strong, 1984). The para-intrusive phases show enrichment or specialization in Sn, W, B, F, Cu, Zn, Au and Ag. The specialized nature of these rocks of the Westfield-Caledonia area is illustrated by the ternary relationships of Sn-Fli and B-F-Li (Fig. 12-4). These plots include data for the entire para-intrusive suite in the study areas up to greisenized granitoid. The well developed greisens of Table 12-2 were not included but are an extension of the trends depicted on Figure 12-4. The dominance of boron in the rocks of this suite, especially the albite-bearing Breccia Matrix and tourmaline-bearing well developed greisens may be significant since fluorine although elevated, does not reach levels expected for rocks of granophile element mineralization in the Meguma Zone.

Chatterjee and Muecke (1972) have defined a granophile element specialization of Li, Be, F, Rb, Cs, U and Sn in the para-intrusive rocks associated with significant Sn-U mineralizations in the SMB. The elemental specialization assemblage of the para-intrusive suite of the Westfield-Caledonia area varies from that described by Chatterjee and Muecke. Notably absent at Westfield are Rb and U. Figure 12-5 compares Rb and K_2O from the rocks of the Westfield area with data of Chatterjee et al. (1983) for specialized and non-specialized granitoid rocks of Nova Scotia. The data from this study fall within or near the field of non-specialized rocks (cogenetic) of Chatterjee et al. The ternary relationship Rb-Ba-Sr also shows the absence of Rb specialization within the Westfield-Caledonia suite (Fig. 12-6). The entire data set remains predominantly within the anomalous granite and normal granite fields of El Bouseilly and El Sokkary (1975). Meguma Zone cogenetic and especially para-intrusive granitoid rocks commonly fall within the strongly differentiated granite field (A on Fig. 12-6;

Table 12-2. Trace element assays for quartz-muscovite greisen and polymetallic sulphide-bearing quartz veins from the Westfield-Caledonia area.

Sample No.	Quartz-muscovite greisen						Greisenized Meguma Group country rock					
	A7C-7507	W-1	W-3	W-4	W-6		A7C-7525	A7C-7559	A7C-7563	A7C-7564	A7C-7565	W-2
As (ppm)	-	-	-	-	-	-	-	-	-	-	-	-
Cu (ppm)	6	70	38	58	20	44	50	130	87	29	90	90
Pb (ppm)	26	20	5	14	20	9	10	18	20	12	70	70
Zn (ppm)	26	120	112	37	26	80	89	57	74	130	185	185
Sn (ppm)	6	7	38	44	75	21	7	120	77	5	59	59
W (ppm)	16	4	16	20	4	4	4	<4	8	4	4	4
Mo (ppm)	7140	9	4	7	4	4	10	6	3	8	4	4
Be (ppm)	2	6	7	14	6	7	3	6	7	4	6	6
Bi (ppm)	-	-	-	-	-	-	-	-	-	-	-	-
Ag (ppm)	0.5	0.1	0.1	0.2	0.2	<0.1	0.1	0.3	<0.1	<0.1	0.1	0.1
Au (ppb)	19	-	-	-	-	-	-	190	-	-	-	-
Quartz veins with polymetallic sulphides ± scheelite ± cassiterite												
Sample No.	A7C-7562	A7C-7566	A7C-7567	A7C-7568	A7C-7569	A7C-7508	A7C-7509	A7C-7510	A7C-7511			
As (ppm)	-	-	-	-	27000	18500	8000	17000	22500	-	-	-
Cu (ppm)	53	4	735	44	26	6	2	2	10	-	-	-
Pb (ppm)	230	16	2	5480	3000	1325	74	840	800	-	-	-
Zn (ppm)	12	9	18	21800	428	276	17	456	42	-	-	-
Sn (ppm)	60	1200	19	6	5	92	10	120	7	-	-	-
W (ppm)	<4	55200	12	<4	<4	<4	<4	12	40	-	-	-
Mo (ppm)	7	10	4	9	7	5	6	5	7	-	-	-
Be (ppm)	3	2	1	2	3	-	-	-	-	-	-	-
Bi (ppm)	-	-	-	-	138	208	15	200	138	-	-	-
Ag (ppm)	11.6	<0.1	0.4	117.0	77	42.0	2.7	27.0	23.0	-	-	-
Au (ppb)	37,190	-	6	160	160	57	20	47	37	-	-	-

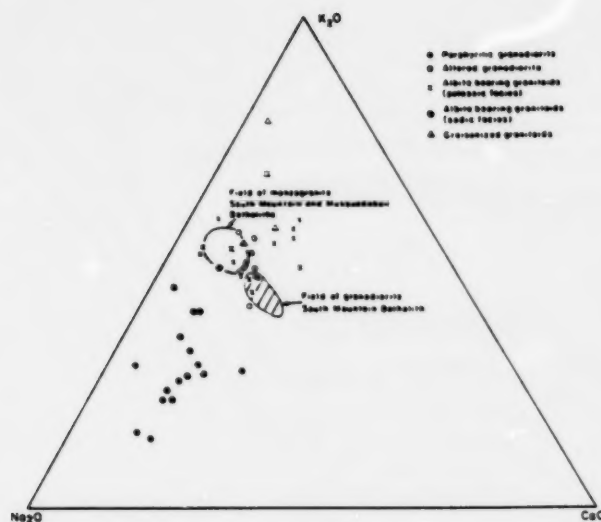


Figure 12-3. Ternary plot of K_2O - Na_2O - CaO for granitoid samples from the Westfield-Caledonia area. Fields of granodiorite and monzogranite are from McKenzie and Clarke (1975), and MacDonald and Clarke (in press).



Figure 12-4. Sn-F-Li and B-F-Li ternary relationships of the granitoid rocks of the Westfield-Caledonia area.

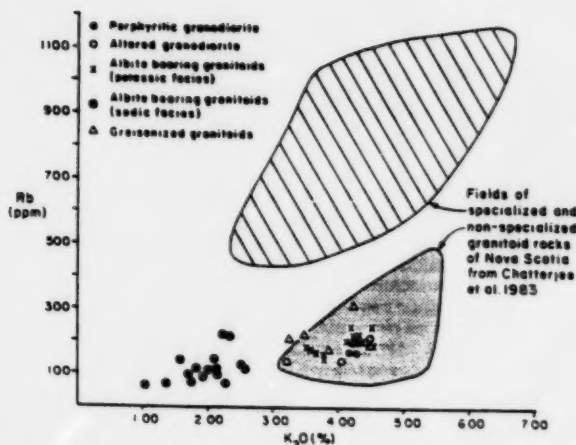


Figure 12-5. Rb-K₂O plot comparing the granitoid rocks of the Westfield-Caledonia area with other Meguma Zone specialized and non-specialized granitoid rocks.

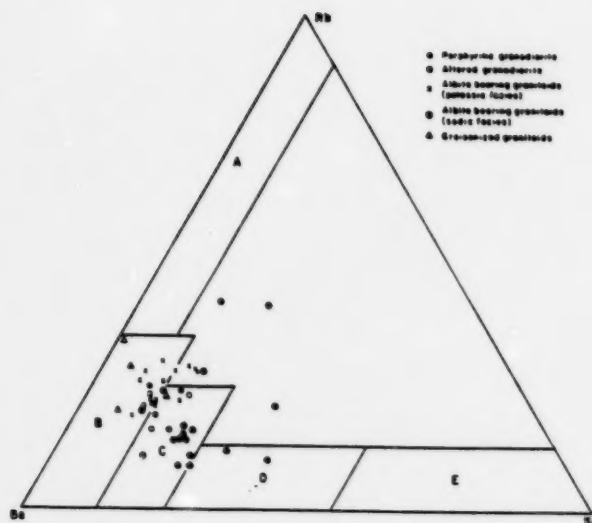


Figure 12-6. Ternary plot of Rb-Ba-Sr for the granitoid rocks of the Westfield-Caledonia area. Fields according to El Bouseilly and El Sokkary (1975) are: A = strongly differentiated granite; B = normal granite; C = anomalous granite; D = granodiorite and quartz diorite; E = diorite.

Chatterjee and Strong, in press). The Rb supposedly reflects the highly evolved character of these host granitoids.

A notable feature of some members of the paraintrusive suite from the Westfield-Caledonia area is a specialization of Au and Ag along with the typically granophile elements. Chatterjee (1983) has noted an association of Ag with significant granite hosted Sn-U deposits of the Meguma Zone. The possibility of specialization of Au within these same paraintrusive rocks has been alluded to in the past but only recently has been investigated. Figure 12-7 is a histogram of gold concentration in samples of both the cogenetic and paraintrusive rocks of the Meguma Zone. The data set in Figure 12-7 includes the data from Table 12-1 and unpublished granitoid rock data of the author and A. K. Chatterjee.

O'Reilly (1984) reported a sample of K-feldspathized porphyry from the Westfield area contained 10 940 ppb Au (sample A7C-7530 in Table 12-1). The sample was split and resubmitted four times and returned a similar result each time. The outcrop was revisited and six samples collected, all of which failed to assay greater than 10 ppb Au. It is felt no cross contamination occurred in sample preparation, as the sample lot record indicated this sample was not prepared with gold-bearing samples. Conceivably the sample contained either one or more small grains of gold, or the sample was prepared in a crusher in which a gold-bearing sample from another lot was crushed. In either case the gold content in this sample does not represent the norm for the K-feldspathized porphyry but is in fact somewhat of a curiosity.

Although the paraintrusive rocks of the Westfield-Caledonia area show elevated Au and Ag content it is the polymetallic sulphide-bearing quartz fissure veins which appear to hold most potential for economic concentrations (Table 12-2). It is in this environment that past exploration has been focused.

DISCUSSION

The granitoid rocks of the Westfield-Caledonia area define a paraintrusive suite associated with granophile and precious metal mineralization. The geochemical signature of the suite shows typical specialization in granophile elements but the absence of Rb and U from the specialization assemblage (Figs. 12-5, 12-6) sets this area apart from other paraintrusive suites described from Meguma Zone granitoids (Chatterjee and Muecke, 1982; Chatterjee and Strong, 1985). The presence of significantly elevated gold levels within the paraintrusive rocks and associated quartz-sulphide veins at Westfield also distinguishes this area from other Meguma Zone examples. The lack of Rb and U specialization and the presence of Au specialization provides clues to the genesis of the paraintrusive rocks of this area if one reviews the literature.

Boyle (1979 p. 295) described gold and silver deposits associated with acidic and basic intrusions in Western Canada, Chile, USSR and Zaire. In these regions the host intrusions are

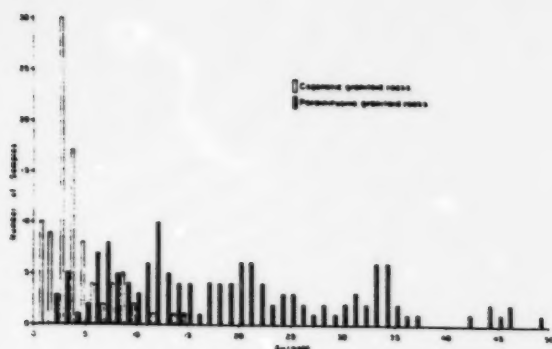


Figure 12-7. Histogram of gold content of Meguma Zone granitoid rocks. All determined by neutron activation analyses.

porphyry dykes and sills, granite-granodiorite stocks and the peripheral phases of granitic batholiths. Boyle also cites alteration processes such as sericitization, silicification, feldspathization, pyritization and carbonitization as associated with the mineralizing episode. The Westfield area shows many of these features. For example, the paraintrusive suite consists of rocks containing various porphyries developed from a granodiorite host at the periphery of a large predominantly granodiorite-monzogranite batholith. Various albited porphyry plugs and dykes have emanated outward from the complex. Alterations associated with formation of the suite included both sodic and potassic feldspathization, pyritization, greisenization and late silicification. Carbonitization within the Westfield-Caledonia suite has been extensive as evidenced by samples of porphyry containing accessory to minor calcite.

Gavrilov and Onikhimovskiy (1983) have noted a close association of Au with Sn, Ag, Cu, Pb, Zn and Mo in Middle Carboniferous sericitized and silicified porphyries with an average Au concentration up to twenty times greater than background. This observation compares with the data of Figure 12-7. They also cited other examples of Sn-Au associations and concluded that Au and Sn, contrary to popular belief, are not elemental antagonists.

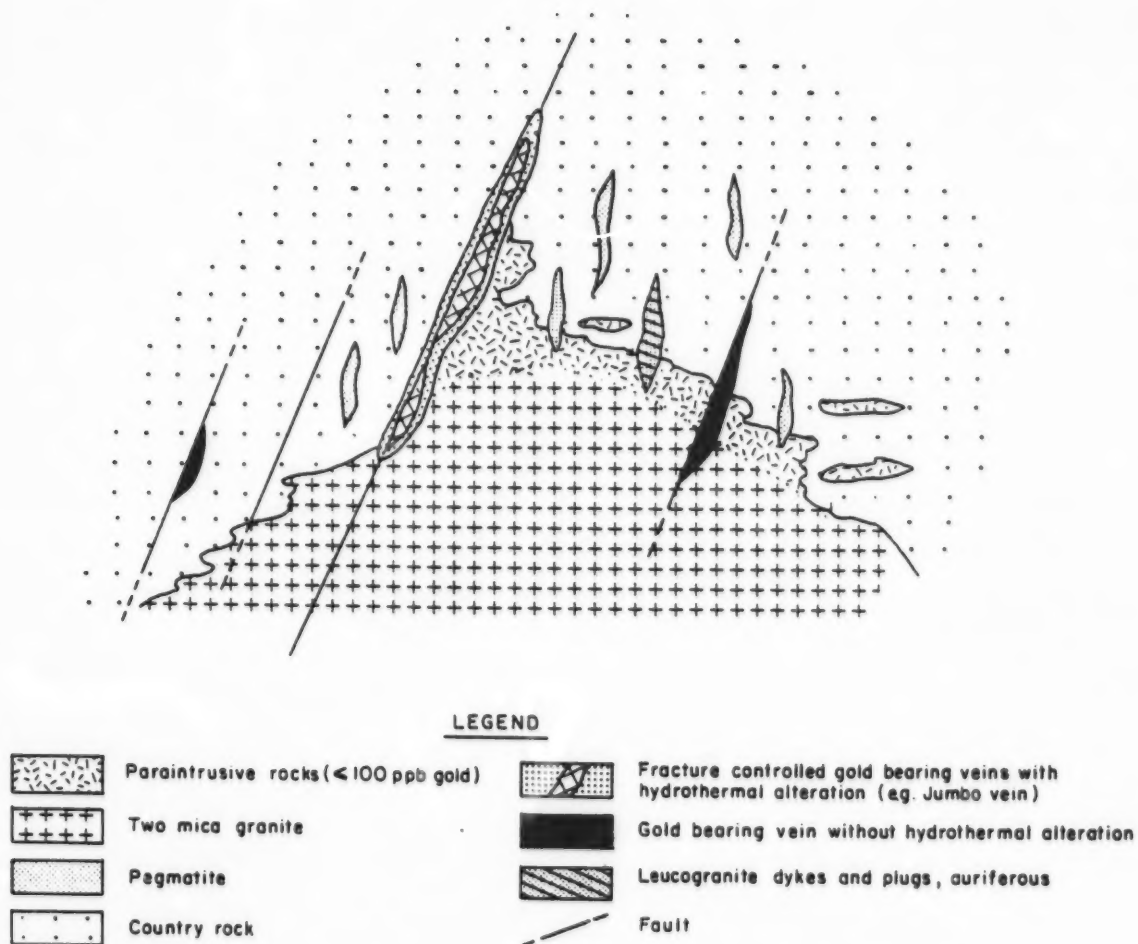
Crouzet et al. (1979) and Ahmadzadeh et al. (1984) discussed the genesis of gold deposits of the Massif Central of France. The similarities of the granitic terranes of Eastern Canada and Western Europe are outlined by Stong (1980). The summary model of Crouzet et al. (1979) for gold deposits of the Massif Central is shown in Figure 12-8 and compares very closely with features observed in the Westfield-Caledonia area. Note the model has been translated to English and modified slightly to meet terminology used in this paper. Features of note are: (1) the association of auriferous and argentiferous polymetallic sulphides with granophile elements; (2) structurally controlled quartz veins (shear and fault zones) and silicified zones hosting

most of the significant precious metal mineralization; (3) the presence of paraintrusive rocks of elevated gold content located in an apical or peripheral portion of a larger complex; (4) the Permo-Carboniferous age of the paraintrusive rocks.

The lack of Rb and U specialization of the Westfield suite is believed to reflect the observation that these rocks were developed from a granodiorite parent magma. This relationship requires that separation of the fluid phase must have occurred from the more primitive granodiorite magma. A fluid phase separated from the magma at this early stage would likely be less enriched in elements such as Rb and U. The paraintrusive rocks described by Chatterjee and Muecke (1982), and Chatterjee and Strong (1985) relate to such areas as East Kemptville, Long Lake and Millet Brook where the generator of the fluid phase was likely leucomonzogranitic or leucogranitic in composition. A fluid phase separated at that point of magmatic differentiation would be much more enriched in late stage elements such as Rb and U.

The presence of anomalous gold in the Westfield-Caledonia suite may also be an indication of a more primitive parent of the fluid phase. Khitrinov and Mel'tser (1979) have found that elevated levels of CO_2 , Cl and Cl of a magma can facilitate early saturation with water allowing separation of fluid at an early stage and at high temperature. Elevated temperature and a high chloride concentration favor the accumulation of gold as chloride complexes in the fluid phase. From this, the anomalous gold content in the paraintrusive rocks at Westfield may be an indicator of a high Cl activity. The presence of accessory and minor calcite within these rocks confirms a high CO_2 activity of the fluid. The coincidence of both elevated CO_2 and Cl within the Westfield system suggests the separation of the fluid phase in the higher temperature regime of the granodiorite could have occurred.

Moiseenko et al. (1970), and Khitrinov and Mel'tser (1979) have noted the association of gold mineralization with more mafic granitoids and also with albitic alteration. The former authors proposed that postmagmatic alteration destroys pre-existing minerals releasing the minute Au content to the fluid phase. If these fluids are receptive to this input of Au ions, i.e. Cl and CO_2 enriched and elevated temperature, then the Au enriched fluids are carried throughout the hydrothermal system. They have also observed that an increase of sodium content in the solutions is also favourable for gold concentration. Within the Westfield-Caledonia paraintrusive suite the petrography and major element chemistry has shown that albitic or sodic alteration predominated the system (Fig. 12-3). It is unclear as to why sodic alteration is dominant within this system but this too may reflect the nature of the fluid phase.



Modified after Crouzet et al., 1979

Figure 12-8. Summary model of gold mineralization associated with the Massif Central of France. Modified to compare terminology used in this paper.

CHAPTER 13

A MAJOR POST-VISEAN SINISTRAL SHEAR ZONE - NEW PERSPECTIVES
ON DEVONIAN AND CARBONIFEROUS ROCKS
OF SOUTHERN NOVA SCOTIA

P. S. GILES

Giles, P. S., A major post-Visean sinistral shear zone - new perspectives on Devonian and Carboniferous Rocks of southern Nova Scotia; in Guide to the Granites and Mineral Deposits of Southwestern Nova Scotia, eds. A. K. Chatterjee and D. B. Clarke; Nova Scotia Department of Mines and Energy, Preprint of Paper 85-3.

ABSTRACT

The Tobiatic Fault Zone is defined as a major zone of sinistral shear extending more than 400 kilometres from the Wedgeport area of extreme southwestern Nova Scotia, northeasterly through the heart of the South Mountain Batholith (SMB) in the New Ross-Vaughan district, into post-Visean fault systems in the Shubenacadie Basin. Reconstruction which removes an estimated 110 kilometres of post-Visean sinistral displacement along the shear zone is supported by geological, geophysical and paleomagnetic data and creates an entirely new perspective from which to view the Devonian and Carboniferous geological history of southern Nova Scotia.

Reconstruction merges the two largest Carboniferous basins of southern Nova Scotia into a single northeasterly trending synclinalorium which functioned as a linear depositional trough in Tournaisian-Visean time. Major peraluminous granitoid batholiths with strong temporal and petrogenetic similarities are combined in the reconstruction to form the S-M-L Complex measuring 180 kilometres by 120 kilometres in dimension and characterized by strong negative Bouguer Gravity expression.

The Tobiatic Fault Zone is spatially related to metallogenic domains characterized by granophile deposits of both intra- and peri-batholithic character, generated by specialized granites of Carboniferous age. The Fault Zone may also link genetically with polymetallic Pb-Zn-Ba deposits distributed along its trace within Carboniferous strata.

Sinistral shear on the Tobiatic Fault Zone which may extend for 750 kilometres along the Appalachian trend, is not yet reconciled with plate tectonic models for Devonian and Carboniferous time which invoke dextral plate motions along the Appalachian-Hercynian fold belt. Tentative rationalization calls for a complex history for the Tobiatic Fault Zone as a long-lived Paleozoic structure.

INTRODUCTION

The importance of major northeast-trending wrench faults to the understanding of the northern Appalachians during Devonian and Carboniferous time was first stressed by Webb (1963, 1969). This paper presents geological and geophysical evidence for a previously unrecognized fault zone in southern Nova Scotia, here termed the Tobiatic Fault Zone (TFZ), on which 110 kilometres of post-Visean, probably Westphalian to Permian sinistral displacement can be shown.

The trace of the Fault Zone explains the presence and orientation of linear steep gravity gradients in parts of southwestern Nova Scotia. Removal of the sinistral displacement realigns major structural/stratigraphic basins of Tournaisian/Visean age and simultaneously "creates" a major plutonic complex comprising the South Mountain and Musquodoboit Batholiths, and the Liscomb Pluton. The reconstruction is supported by the merger of regional Bouguer Gravity lows and radiometric highs. The fault

zone rationalizes the distribution of metallogenic domains characterized by granophile mineral deposits and may be genetically significant in explaining the distribution of other types of mineral deposits along its trace.

Significant sinistral fault movements of Late Carboniferous age on the TFZ suggest comparison with the Harvey-Hopewell Fault of southern New Brunswick (Webb, 1963) but are difficult to reconcile with post-Westphalian "A" dextral movements on the Cobequid-Chedabucto Fault (Eisbacher, 1969, 1970) with which the TFZ appears to merge at its eastern extremities. Like many major fault systems in Atlantic Canada, the Tobiatic Fault Zone may have a complex history involving both dextral and sinistral movement (Webb, 1963, 1969; Keppie, 1982).

GEOLOGICAL SETTING OF SOUTHERN NOVA SCOTIA

Southern Nova Scotia is dominated by flyschoid rocks of the Cambro-Ordovician Meguma Group, deformed during the Acadian Orogeny into a series of linear to slightly arcuate major folds trending generally northeasterly. This thick succession comprises only two formations, a lower Goldenville Formation characterized by variably bedded meta-greywacke, and an upper Halifax Formation dominated by slate and siltstone. Various aspects of Meguma Group stratigraphy and sedimentology have been described by Schenk (1970), Harris (1971, 1975), and Harris and Schenk (1975).

Other Lower Paleozoic rock units in southern Nova Scotia include the White Rock, Kentville, New Canada and Torbrook Formations. Summary descriptions of each of these formations are available in Keppie (1977) and no further attempt to describe the rocks is made in this paper. Like the underlying Meguma Group, each of these units was deformed by the Acadian Orogeny.

Subsequent to the major deformational events of the Acadian, southern Nova Scotia was subjected to igneous intrusion of peraluminous granitoid rocks today represented by several plutonic bodies, the largest of which is the SMB (Fig. 13-1A). This major composite Batholith has been the focus of much recent research and has been extensively described in the literature (McKenzie and Clarke, 1975; Clarke et al., 1976; Smith and Turek, 1976; Clarke and Halliday, 1980; Muecke and Clarke, 1981; Reynolds et al., 1981; Clarke and Chatterjee, 1985; Clarke and Muecke, 1985).

Aspects of this Batholith which have also been emphasized are its relationship to polymetallic mineral deposits (Chatterjee, 1980; Chatterjee et al., 1982; O'Reilly et al., 1982; Chatterjee, 1983; Chatterjee et al., 1983; Chatterjee and Strong, 1984A and B; 1985) the most significant of which is the East Kemptville Sn-Cu-Zn-Ag mine of Rio Algom.

Recent emphasis has also included the Musquodoboit Batholith (MB) (MacDonald and Clarke, 1983) which is, in part, highly similar to the SMB and shares a common petrogenetic and metallogenic history with that body.

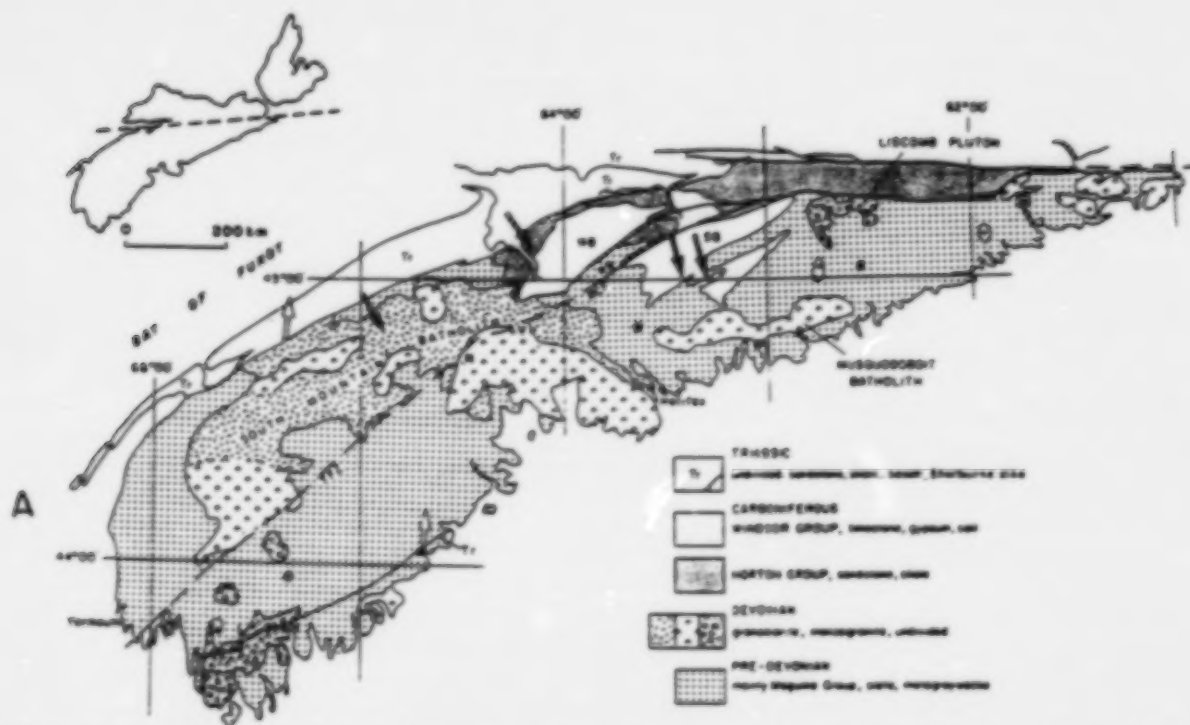


Figure 13-1A: Simplified geological map of southern Nova Scotia, modified slightly from Keppie (1979), showing the distribution of Lower Paleozoic rocks, granitoid rocks of Devonian to Carboniferous age, and Carboniferous rocks of Tournaisian and Viséan age. The heavy line marks the trace of the Tectonic Fault Zone (TFZ). Major igneous bodies discussed in the text are indicated. WB = Windsor Basin; SB = Shubenacadie Basin; RB = Rawdon Block.

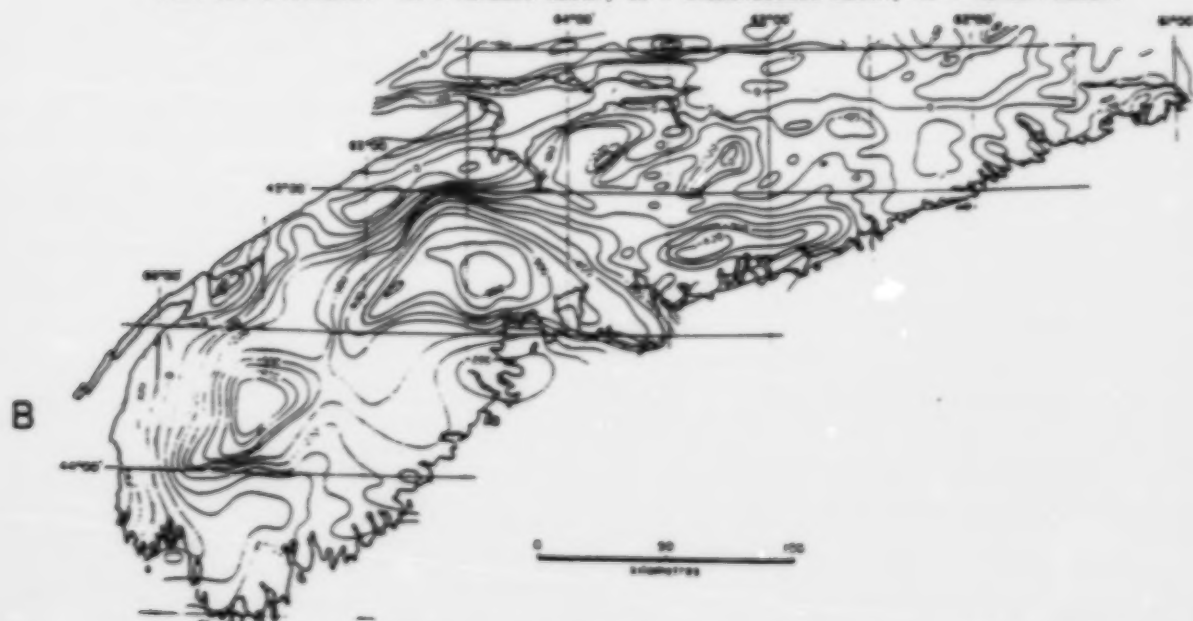


Figure 13-1B: Bouguer Gravity map of southern Nova Scotia, after Keppie (1979); note linear zones of steep gradient in southwestern Nova Scotia and prominent gravity lows over major granitoid bodies.

In configuration, the SMB forms an arcuate, "banana-shaped" body atypical of large granite batholiths elsewhere (Pitcher, 1978). When considered with the contiguous Musquodoboit Batholith and Liacomb Pluton, these major bodies present a broad Z-shaped configuration. These three bodies are an integral part of this paper which deals with changes in their present configuration. The age of the SMB and MB lies between 372 and 360 Ma (Reynolds et al., 1981).

The Musquodoboit, Shubenacadie and Windsor Basins (Fig. 13-1A, 13-2), together with the so-called St. Marys graben adjacent to the Cobequid-Chedabucto Fault System of Webb (1969; Fig. 13-1A), define major outcrop areas of the Tournaisian and Viséan Horton and Windsor Groups. These major basins appear to represent not only present-day structural features, but also Carboniferous depositional features.

Descriptions of the Horton Group are found in Bell (1929, 1960) and in numerous unpublished graduate theses completed at Acadia University in Wolfville. In the northeastern parts of the Windsor Basin, the Horton may reach 1000 metres in thickness, and consists of a lower Horton Bluff Formation (grey sandstone, grey shale) and upper Cheverie Formation, characterized by red sandstone and shales.

On the Rawdon Block (Fig. 13-1A), the Horton Group is thick and seemingly complete. However, within the immediately contiguous Shubenacadie Basin to the southeast, the Horton Group is represented by less than 100 metres of red sandstones, conglomerates and shales assigned by Giles and Boehner (1981) to the Coldstream Formation, and correlated with the Cheverie Formation of the Windsor Basin.

Descriptions of the Windsor Group in southern Nova Scotia which consists of major deposits of evaporites with rhythmic intercalations of marine limestone and terrestrial red and green siltstones, are found in Bell (1929), Moore (1967), Giles and Boehner (1979), Giles (1982), Boehner (1984), Howie (1984) and Schenk (1984). Moore (1967) first demonstrated the close similarity of Windsor strata between the Shubenacadie and Windsor Basins. This observation is in complete agreement with the writer's personal assessment of these two major basins.

Major northeasterly and east-west trending fault systems cut the Carboniferous strata. The continuation of many of these structures into the Meguma strata has been confirmed, although recognition of major northeasterly trending structures in Meguma outcrop areas is hampered by the strong northeasterly structural grain and a general absence of recognizable crosscutting relationships.

Deformation of Carboniferous strata ranges considerably in intensity and seems to be most intense near major fault traces. In the plastic evaporitic strata of the Windsor Group, this deformation results locally in extremely complicated fold patterns with associated thrusts or decollement structures (or both). This deformation is considered by Kappas (1982) to

represent the Hercynian Orogeny in southern Nova Scotia. Late Westphalian strata appear to be little affected by this event.

DEFINITION OF THE TOBIATIC FAULT ZONE

The Tobiatric Fault Zone is defined in this study as a major, previously unrecognized zone of sinistral fault movement, extending southwesterly from known faults cutting latest Viséan strata of the Shubenacadie Basin, through the New Ross-Vaughan district, at least to the Yarmouth-Wedgeport area of extreme southwestern Nova Scotia. The Rushmere Lake Shear Zone of Smith (1985) and unnamed shear zones truncating the Wedgeport Pluton in southwestern Nova Scotia are defined as parts of the Tobiatric Fault Zone.

SHUBENACADIE BASIN FAULT SYSTEMS

Major northeasterly trending faults mapped by Giles and Boehner (1979, 1981) in the Shubenacadie Basin (Fig. 13-2) include the Rawdon, Raulston Corner, Brentwood-Graham Hill and Otter Brook Faults which collectively cut latest Viséan strata, and define the northwestern and northern limits of the Shubenacadie Basin. The Meadowvale and Milford Station Faults are parallel trending faults which follow the southern boundary of the basin and define it as a true graben (Boehner, 1984; Howie, 1984). At their northeasterly limits, these faults are either truncated by the West River St. Marys Fault of Benson (1974) or may merge with that Fault.

The Otter Brook Fault, extending southwesterly to merge with the Raulston Corner Fault beneath the Salem Thrust, appears to coincide with major changes in Windsor Group thickness and coincides with a major change in the stratigraphic assemblage comprising the Horton Group, particularly apparent in comparing the Horton succession within the Shubenacadie Basin to that preserved on the Rawdon Block (Fig. 13-2). Otter Brook and Raulston Corner Faults are here defined as part of the Tobiatric Fault Zone (TFZ).

Youngest strata cut by the TFZ in the Shubenacadie Basin are assigned to the Watering Brook Formation (Giles and Boehner, 1981) dated by Utting (1980) as latest Viséan in age. Fault movements younger than latest Viséan cannot be precluded, since Westphalian strata in the same basin nowhere lie on the trace of the TFZ, and may therefore have escaped fault-related deformation.

VAUGHAN-NEW ROSS DISTRICT

Extension of the Tobiatric Fault Zone through the Vaughan-New Ross district on the projected southwesterly trend of the Raulston Corner Fault is supported by direct evidence of shear deformation, by the abrupt termination of major rock units on sublinear boundaries, and by the configuration of regional gravity anomalies in that area.

The projected trace of the Tobiatric Fault Zone follows the northwesterly limit of a major monzogranite body extending from Halifax to the

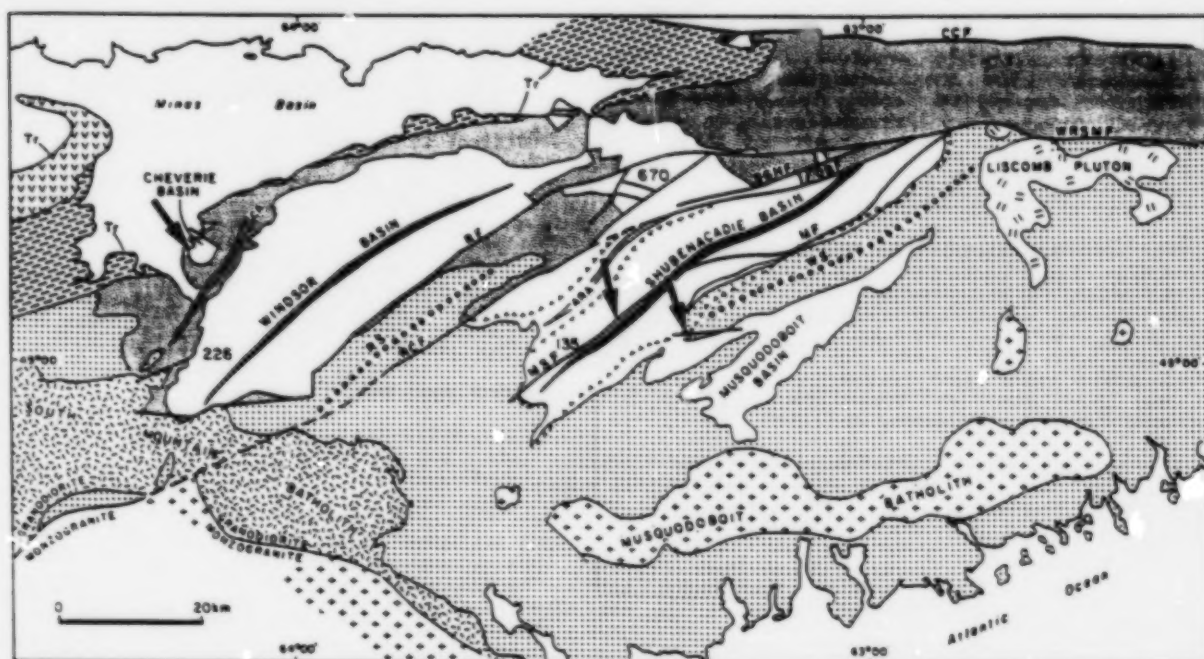


Figure 13-2: Major geological features used to constrain the reconstruction of Figure 13-4a. Axial zones of major Carboniferous basins are shown in black. RF = Rawdon Fault; RCF = Raulston Corner Fault; MF = Meadowvale Fault; MSF = Milford Station Fault; WRSF = West River St. Marys Fault; CCF = Cobequid-Chedabucto Fault; ARA = Admiral Rock Anticline; CA = Cheverie Anticline; RS = Rawdon Synclinorium; WS = Wittenberg Synclinorium; solid arrows = Carboniferous paleomagnetic declination (see text); numbers refer to average thickness, in metres, of the Green Oaks Formation of the Upper Windsor Group; hatched line shows limits of Coldstream Formation beneath the Windsor Group; v = Triassic basalt, dashed pattern Triassic sedimentary rocks; other notation as in Figure 13-1a.

New Ross-Vaughan district and terminating against granodiorite or against rocks of two Meguma "inliers". Turner (1981) reported complex faulting coinciding with marked VLF-EM anomalies along the northeasterly trending contact between the larger Meguma inlier (Fig. 13-2) and the monzogranite body to the south. O'Reilly (1982) noted numerous faults and shear breccias in the monzogranite near the boundary stretching between Vaughan and New Ross (Fig. 13-1A), a distance of approximately 20 kilometres. Identification of this major geological boundary as a northeasterly trending fault zone also provides a reasonable explanation for abrupt orthogonal termination of the monzogranite-granodiorite contact extending from Halifax to Vaughan (Fig. 13-2).

The major elongate Bouguer Gravity low which extends from Halifax northwesterly and is centred south of the New Ross-Vaughan district, deflects sharply in trend on the regional anomaly map to westerly and southwesterly near the postulated trace of the fault zone (Fig. 13-1B). Re-contouring of the regional data at five milligal intervals reveals that the southwesterly extension of this large anomaly is probably a discrete but separate low. The postulated trace of the fault zone would separate this low spatially from the major low, and explain the northwesterly

termination of the larger anomaly in the New Ross-Vaughan district.

DOCUMENTATION IN SOUTHWESTERN NOVA SCOTIA

Southwesterly between New Ross and Yarmouth (Fig. 13-1A), two major segments of the southeastern boundary of the SMB are notably linear, closely reflecting the northeasterly structural grain of the Meguma Zone. The implied structural control of these segments of the intrusive boundary is a marked departure from the norm in southern Nova Scotia where granitoid bodies typically cut major folds and are irregularly lobate in configuration (Fig. 13-1A). Evidence now exists to suggest that the largest of these linear segments, stretching from the southwestern tip of the SMB northeasterly to the Tobistic Game Sanctuary northwest of Lake Rossignol, is controlled in large part by a zone of mylonites and fluid breccias currently under investigation by Smith (in press). This zone of intense shear ranges in width from a few tens of metres to several hundred metres, dips steeply and has been traced for more than ten kilometres by diamond drilling. Shear deformation of country rock extends for more than a kilometre from the centre of the mylonite zone. For at least half of its length, the shear zone follows

the granitoid-Meguma contact with clear evidence of shear deformation of the granitoid.

At the northwestern extremity of this linear segment of the granitoid boundary, A. K. Chatterjee (personal communication, 1985) has observed intense shear brecciation in zones cutting the SMB, thereby suggesting the extension of the zone of intense shear of Smith more than 35 kilometres in a northeasterly direction. Steep gradients and strong northeasterly trends in gravity contours (Fig. 13-18) adjacent to this segment of the Batholith boundary in southwestern Nova Scotia, support the presence of a steeply dipping fault zone as postulated.

To the southwest, on the projection of the two major linear segments of the South Mountain Batholith-Meguma boundary, Chatterjee and Keppie (1982) mapped northeasterly-trending shear zones cutting through the northern-most part of the Wedgeport Pluton south of Yarmouth, dated at 316 ± 17 Ma, (Rb/Sr whole rock) and 313 ± 3 Ma (concordant zircon age; Keppie et al., 1983). The northeasterly trend of these shear zones suggests that they may represent part of the same major zone described by Smith. In the Wedgeport-Yarmouth area, contours of regional Bouguer gravity deflect sharply from northerly to southwesterly where they intersect the shear zones mapped by Chatterjee and Keppie (1982) (Fig. 13-18). Deformation of the Wedgeport Pluton clearly indicates shear movement of Westphalian or younger age.

SOUTHWESTERLY EXTENSION

In the Gulf of Maine, Ballard and Uchupi (1975) delineated through the study of seismic-reflection profiles, a major fault system termed the Georges Basin Fault System (GBFS) and a parallel, unnamed fault system (Fig. 13-3), at least 375 kilometres in combined length, which project reasonably into onshore Nova Scotia through the Wedgeport area. These fault systems were interpreted as normal faults, upthrown on the north, which controlled Triassic sedimentation. Ballard and Uchupi postulated lateral movement on the GBFS and parallel structures coincident with the subsequent separation of North America and Africa in Late Triassic-Early Jurassic time.

In summary, major fault systems of post-Visean age can be reasonably extended southwesterly as far as Yarmouth using fragmentary but convincing evidence of variably intense shear deformation in a linear zone approaching in dimension that of the Cobequid-Chedabucto Fault System of Webb (1969). Possible continuation into the Gulf of Maine via the Georges Basin Fault System of Ballard and Uchupi (1975) would extend the minimum strike length of the zone to more than 700 kilometres. For convenient reference, this major shear zone is here termed the Tobiatic Fault Zone after the area providing the best physical evidence of intense northeasterly shear.

DEVONIAN-CARBONIFEROUS RECONSTRUCTION

The mis-match of Horton Group formations across the TFZ along the northwestern margin of the

Shubenacadie Basin suggests possible strike-slip along the zone. Closer similarity of Horton-Windsor sequences between the Shubenacadie setting and that known in the southwestern extremities of the Windsor Basin, indicates sinistral offset of the Shubenacadie Basin.

Two fundamental constraints govern the amount of offset which can be postulated. Firstly, the close similarity of Windsor Group strata in the Windsor and Shubenacadie Basins, between which carbonate members less than ten metres thick are correlative and virtually identical (Moore, 1967) although presently separated by more than 50 kilometres, dictates close paleogeographic connection between these two basins. Secondly, any displacement of that part of the SMB lying southeast of the Tobiatic Fault Zone which removes direct contact with the larger part of the Batholith, is not acceptable.

Major structural features of Devonian and Carboniferous age truncated by the TFZ which aid in reconstruction include the major synclinal axis of the Shubenacadie Basin, a parallel anticlinal axis represented by the Admiral Rock Anticline of Giles and Boegner (1981), and the northwesterly trending mapped boundary between monzogranite and granodiorite extending from Halifax to the New Ross area (Fig. 13-2). An additional stratigraphic and structural linear which might support a pre-Carboniferous restoration, is the major synclinal axis which defines the axis of the Rawdon Hills (Fig. 13-2) which is truncated at its southwesterly limits by the trace of the TFZ.

The reconstruction of Figure 13-4A simultaneously restores all of these major features. The Shubenacadie Basin axis becomes a projection of the Windsor Basin axis. The Admiral Rock Anticline projects northeasterly into a similar feature separating the subcircular Cheverie Basin from the Windsor Basin. At its southwestern limit, this anticline projects on the trend of an elongate small pre-Carboniferous ridge of Meguma rocks, suggesting that, like the Admiral Rock Anticline, it is an expression of pre-Carboniferous topography.

The monzogranite-granodiorite boundary, trending at a high angle to the TFZ, is continued across the Fault Zone in the southwestern part of the SMB, using a similar boundary mapped during regional exploration by Kidd Creek Mines Limited, (D. Duncan, personal communication, 1985). The Rawdon Synclinal axis projects southwesterly to continue into the Wittenberg Synclinal axis in the restoration. Thus, the reconstruction appears to satisfy all constraints and link major structural/stratigraphic features of differing age. As a consequence, it seems applicable to a considerable span of Devonian and Carboniferous time and suggests approximately 110 kilometres of post-latest Visean sinistral offset along the TFZ.

The reconstruction presents a new perspective, not only on the configuration of major Carboniferous basins, but on the geometry of granitoid bodies which share a common metallogenic and petrogenetic history. The

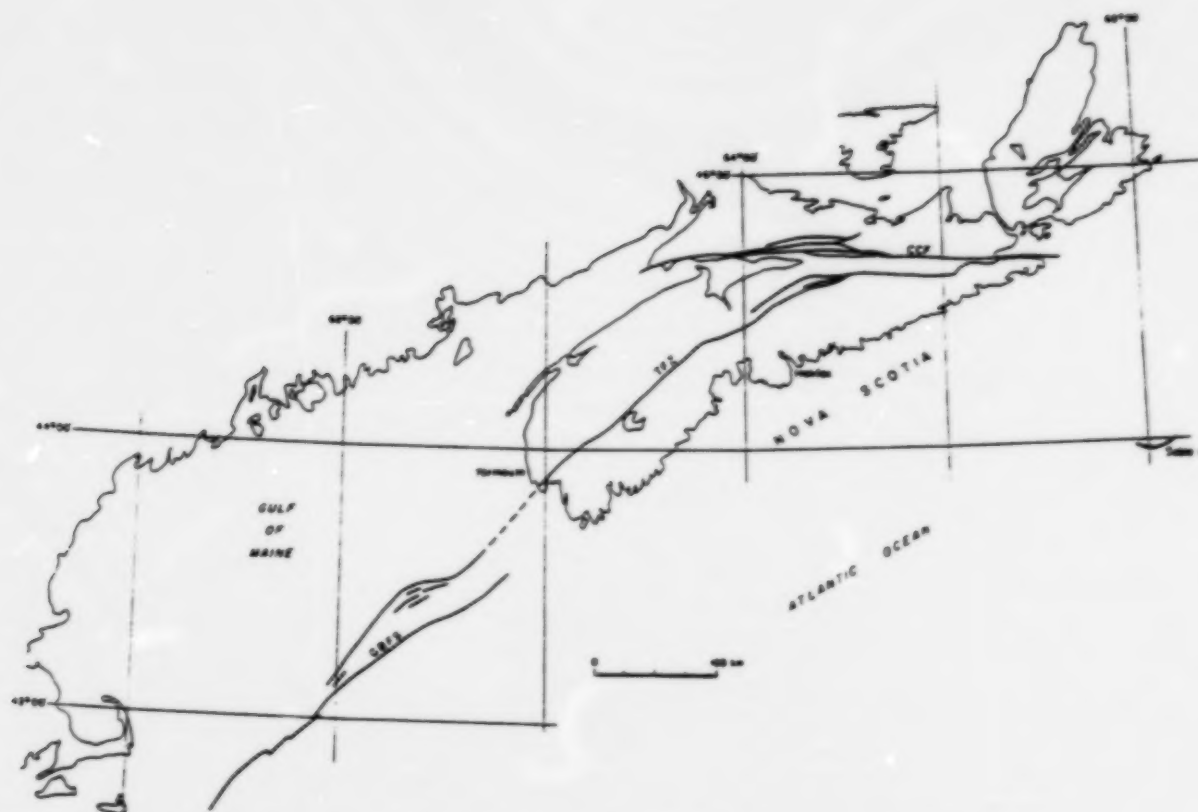


Figure 13-3: Possible extent of the Tobistic Fault Zone (TFZ), with inferred continuation into the Georges Basin Fault System (GBFS) of Ballard and Uchupi (1975) in the Gulf of Maine. The onshore extent of the Cobequid-Chedabucto Fault system (CCF) is shown for comparison.

restored SMB, when considered with the MB and the Liscomb Pluton, is transformed from an enigmatic "banana-shaped" body, to become the principal component of an elliptical plutonic complex approximately 180 kilometres in maximum dimension (Fig. 13-4A). In this restored context, continuation of the Musquodoboit Batholith (M) westerly to link with the South Mountain Batholith (S) beneath a cover of Meguma strata, as argued by MacDonald and Clarke (1985) receives strong support. This concept is here broadened to propose similar continuity northerly into the Liscomb Pluton (L), with potential expression at surface by two small unnamed plutons which help to define the southeasterly part of the major elliptical plutonic complex. This complex is termed, for ease of reference, the S-M-L Complex for its three largest plutonic bodies.

TESTING THE RECONSTRUCTION

BOUGUER GRAVITY

Contours of Bouguer Gravity in Figure 13-4B were transcribed directly from the Geophysical Edition

of the Geological Map of Nova Scotia of Keppie (1979). The more proper approach which would re-contour areas in close proximity to the trace of the Tobistic Fault Zone would clearly lead to charges of bias in interpretation, and would not materially affect the broad gravity configuration.

In the reconstruction of Figure 13-4B, major Bouguer Gravity lows are neatly circumscribed by the outline of the S-M-L complex. The two most pronounced lows are juxtaposed at the southwestern end of the S-M-L ellipse, suggesting former continuity across the zone of a major linear or slightly arcuate Bouguer Gravity low. The perimeter of the ellipse is marked by a discontinuous series of closures relatively high in Bouguer Gravity.

The supporting factors noted above must be counterbalanced by the admission that the reconstruction creates an unacceptable truncation of gravity contours coincident with the trace of the Tobistic Fault Zone at the northeastern end of the S-M-L ellipse (Fig. 13-4B). A suggested

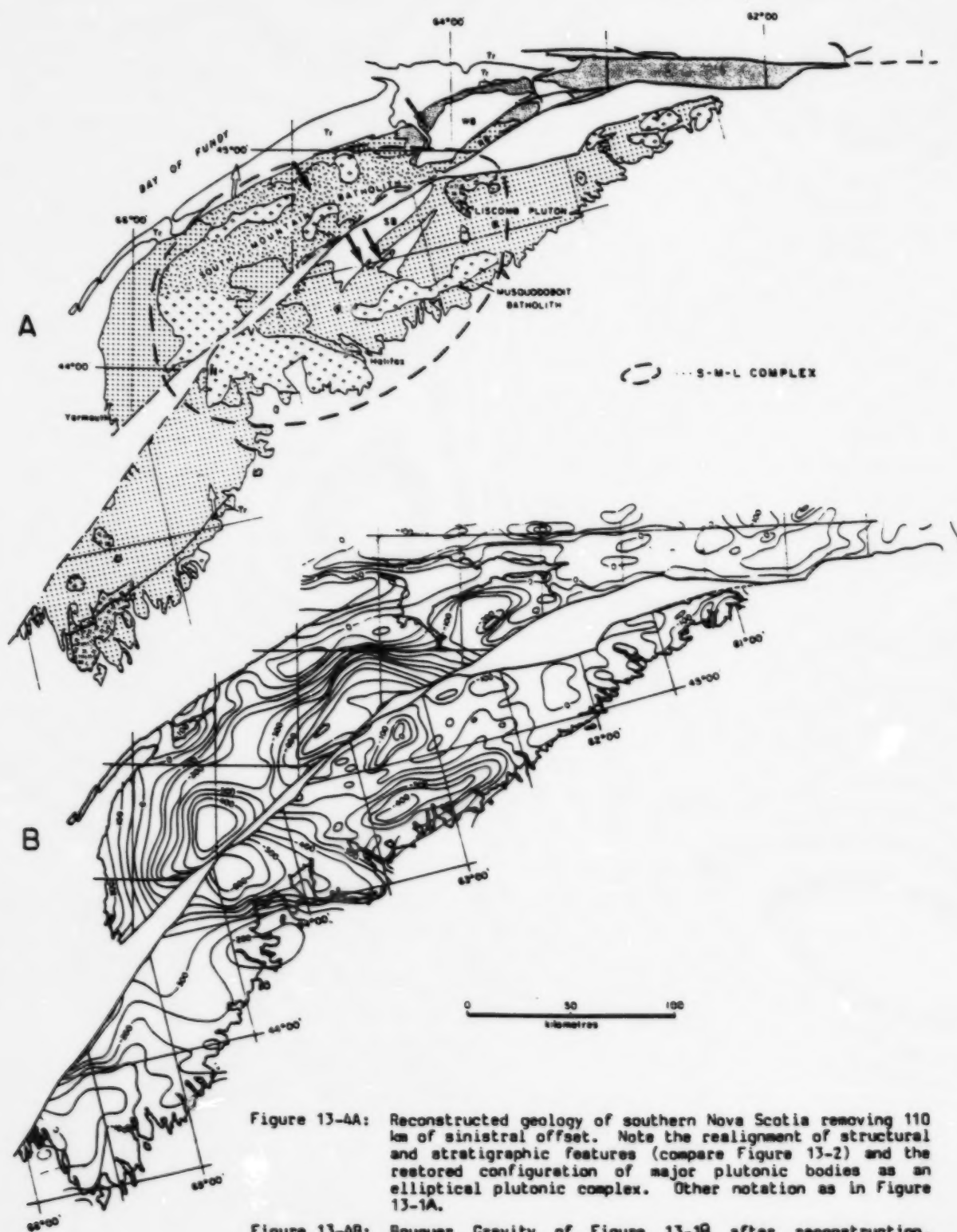


Figure 13-4A: Reconstructed geology of southern Nova Scotia removing 110 km of sinistral offset. Note the realignment of structural and stratigraphic features (compare Figure 13-2) and the restored configuration of major plutonic bodies as an elliptical plutonic complex. Other notation as in Figure 13-1A.

Figure 13-4B: Bouguer Gravity of Figure 13-1B after reconstruction, emphasizing the merging of major anomalies and concentration of regional low gravity expression within the circumscribed S-M-L complex of Figure 13-4A.

explanation for this artifact of the reconstruction is differential vertical movements across the TFZ, down to the southeast, resulting in the preferential preservation of cover rocks (more positive Bouguer Gravity response) southeast of the TFZ. Thus, this termination of gravity contours might reflect differential unroofing of the S-M-L complex, accomplished very soon after cooling, and may be tentatively explained in the reconstruction as a pre-Carboniferous expression of Bouguer Gravity contrast.

AIRBORNE RADIMETRICS

The utility of airborne radiometric data in delineating highly differentiated as well as highly altered para-intrusive rocks of the SMB has been shown by Chatterjee and Muecke (1982). Expressed in terms of equivalent uranium, thorium, potassium, or ratios of these parameters, the surveys note large areas of anomaly centred principally on the New Ross area, extending southeasterly towards Halifax, and on the area of the Davis Lake Pluton, situated near the southwestern extremity of the SMB (Fig. 13-5A).

The reconstruction of Figure 13-4A, when applied to the radiometric data exemplified by the ratio of equivalent uranium to equivalent thorium (Fig. 13-5A and B), merges these two major anomalous areas at the southwestern end of the S-M-L Complex analogous to the merging of major Bouguer Gravity lows. This suggests a regional radioelement distribution linked genetically to the emplacement of the complex in its restored configuration, consistent with uranium enrichment and thorium depletion in comagmatic series with increasing differentiation (Chatterjee and Muecke, 1982).

THE LISCOMB PLUTON

The reconstruction (Fig. 13-4A) suggests former close proximity of the Liscomb Pluton to eastern parts of the SMB, implying petrogenetic similarities of the granitic rocks between the plutonic bodies. Geological mapping of the western half of the Liscomb Pluton combined with limited petrographic data indeed suggest that the evolution of the granitoid rocks are similar. The Liscomb Pluton, like the SMB, is characterized by biotite granodiorite, biotite monzogranite, leucocratic muscovite-biotite monzogranite and late stage leucocratic rocks which include aplites and porphyries.

The Liscomb Pluton is disharmonic with its surroundings implying a posttectonic emplacement and the presence of country rock xenoliths showing various stages of assimilation in the marginal granitic phase, late stage rocks with granophytic textures,miarolitic cavities and quartz-tourmaline veins all point to a high level emplacement environment with perhaps a significant role of volatiles.

In outcrop the granodiorite is grey, medium- to coarse-grained and consists of 45 to 55 per cent plagioclase, 20 to 25 per cent quartz, 5 to 15 per cent K-feldspar and 5 to 10 per cent biotite. Accessory minerals are apatite, zircon,

garnet, and rarely muscovite and chlorite as secondary minerals. Rarely, in a few samples the rock is devoid of K-feldspar and the rock type appears to be quartz-diorite. Individual plagioclase crystals are commonly one to eight millimetres across, though some are as large as two centimetres, and range in composition from An_{18} on the rim to An_{45} in the cores of the more calcic grains. Many crystals show well formed crystal outlines, and oscillatory zoning is common except in very small grains or where zoning has been destroyed by twinning. A small amount of albite forms discontinuous, thin clear rims on microcline and appears to be the result of exsolution. Myrmekite is also common in sections containing considerable microcline, especially on the rims of plagioclase crystals bordering microcline. Biotite occurs as well formed crystals as much as five millimetres across containing abundant inclusions of zircon and apatite. Of the accessory minerals, garnet is important in its occurrence as discrete crystals in biotite clusters and also as clear euhedral crystals in plagioclase feldspar, probably indicative of xenocrystic or primary magmatic origin. It also indicates the peraluminous nature of the granodiorite which is a feature well established for the granodiorite of the SMB.

The biotite monzogranite consists of 40 to 45 per cent quartz, 15 to 25 per cent oligoclase, 20 to 25 per cent K-feldspar and less than 10 per cent accessory biotite, zircon, apatite, tourmaline, garnet, pyrite and muscovite. Oligoclase occurs as subhedral to euhedral crystals having an average composition of about An_{24} . Oscillatory and patchy zoning is characteristic of most grains and some have rims of normally zoned material grading into albite on the outer edge. Quartz either forms inequigranular grains that replace oligoclase, or it occurs interstitially. However, most of the quartz is strained. Irregular shreds of biotite as much as three millimetres across are pleochroic from light yellowish-brown to dark brown and zircon occurs as inclusions where it forms distinct pleochroic halos.

The leucomonzogranite, aplite, porphyries and quartz-garnet-tourmaline veins are all grouped as late stage rocks and these commonly occur as lenticular or tabular bodies and range in thickness from less than a foot thick to large mappable units. The assemblage and abundance of minerals in the late stage rocks differ slightly from those of biotite-monzogranite but muscovite is distinctly a primary phase together with garnet and tourmaline. Likewise, apatite and zircons are present in only some of the rocks and are not consistently found together. Throughout these rocks, the plagioclase is either albite or oligoclase. Biotite occurs either as widely scattered chloritized flakes or as unaltered inclusions in quartz. Much of the muscovite in these rocks may be secondary and occurs in close association with calcite, iron oxide and chlorite.

In summary, the Liscomb Pluton indeed shows important similarities to the SMB, not only in its emplacement history but in its relatively

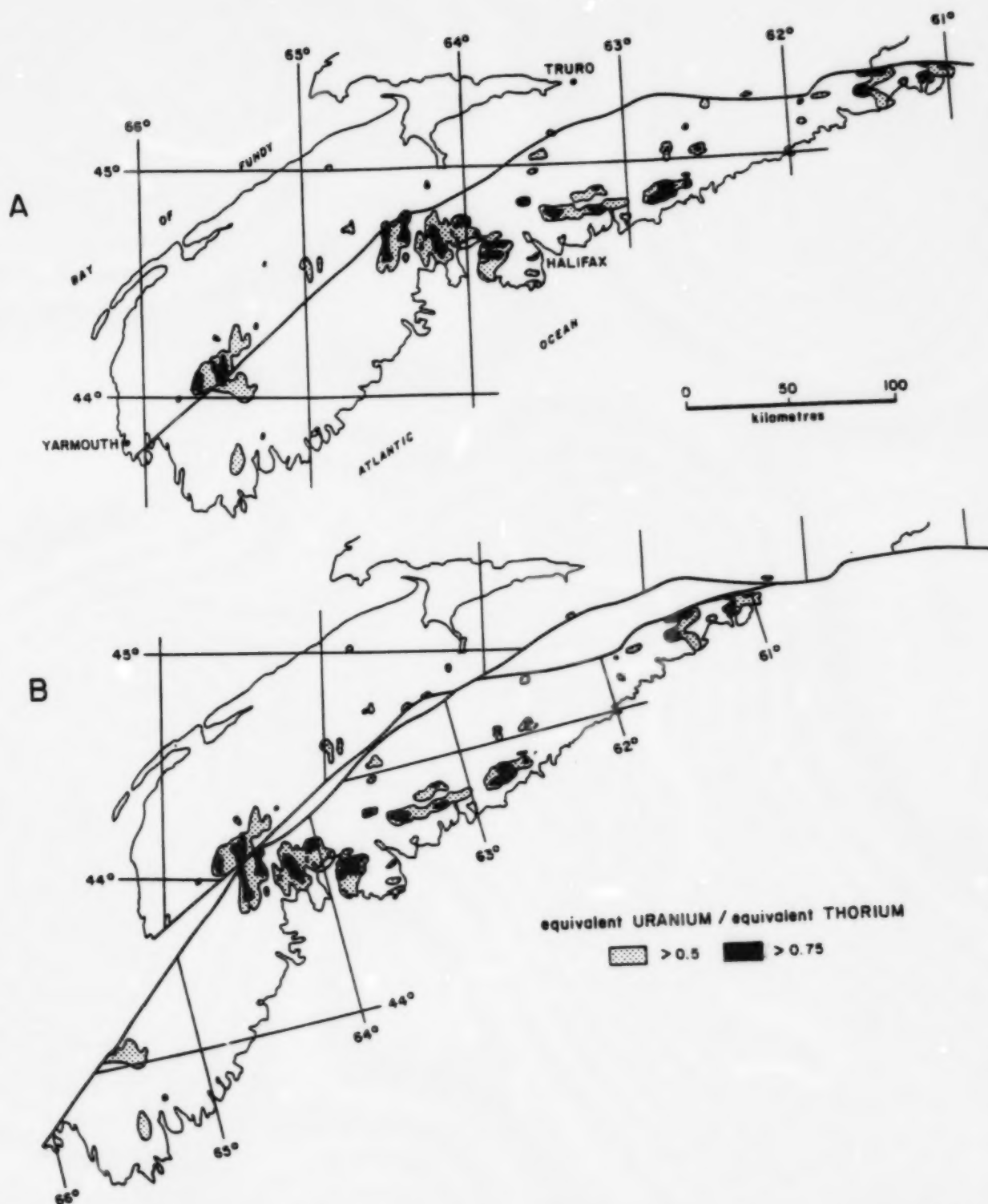


Figure 13-5: Airborne radiometric map (eU/eTh) before (A) and after (B) removal of sinistral displacement along the Tobiatic Fault Zone showing the merger in (b) of major anomalous areas in the southwestern part of the S-M-L complex. Data simplified from an unpublished experimental compilation by the Geological Survey of Canada, published with permission of the Survey.

broad range of granitoid constituents. In this respect, it is important to note the contrast between the Liscomb Pluton and the Musquodoboit Batholith which lacks the megacrystic granodiorite (MacDonald and Clarke, 1985) so characteristic of both the SMB and the Liscomb Pluton. The restored proximity of the Liscomb Pluton to the SMB (Fig. 13-4A) helps to explain their similarity, and in turn suggests that they may share a common mineral potential.

CARBONIFEROUS STRATIGRAPHY AND SEDIMENTOLOGY

The reconstruction of Figure 13-4A implies the merger of the Shubenacadie and Windsor Basins into a single, elongate, northeast-trending synclinal trough which controlled patterns of sedimentation and the resultant facies configuration. Several lines of evidence are compatible with this model.

The Horton Group in the Shubenacadie Basin is represented by the relatively thin (<100 m) Coldstream Formation. Depositional limits, established through detailed mapping and subsurface investigations (Giles and Bohner, 1979, 1981), closely reflect the present basin geometry and define a linear pre-Windsor Valley trending northeasterly and truncated by the major faults of the Tobiatic Fault Zone (Fig. 13-2).

Deep subsurface data are not available to determine major changes in character or thickness of Horton sediments towards the northeast, but it is known that at depths as great as 400 metres, Horton Group rocks are absent near buried pre-Carboniferous ridges represented by the Admiral Rock Anticline of Figure 13-2, in positions well removed from the present basin margins.

In the southwestern extremities of the Windsor Basin, the earlier Horton Bluff Formation thins dramatically and is overstepped by the Cheverie, typically less than 150 metres thick in that area and equivalent to the Coldstream Formation of the Shubenacadie Basin. In the reconstruction, this relationship would be predicted, based on the restored proximity of the Shubenacadie Basin to the southwest.

Patterns of Windsor Group sedimentation indicated by detailed isopachs of carbonate member thickness show that the Windsor Basin was a u-shaped trough opening to the northeast (Moore, 1967). Similar data for the Shubenacadie Basin, generated by research of the writer, suggest that it too was a U-shaped trough, opening to the northeast, but with a depressed basin centre indicated by closure of isopach contours. Bohner (1984) showed that the centre of the Shubenacadie Basin functioned to concentrate saline brines during late Windsor regressions, but did not appear to affect sedimentation during transgressive maxima.

Comparison of formation thickness provides important support for the reconstruction. Using the Green Oaks Formation as defined in the Shubenacadie Basin, average thicknesses can be compared from southwest to northeast along the reconstructed sedimentary trough (Fig. 13-4A), and are suggestive of a northeasterly trend in

thickness increase compatible with a major depositional feature opening to the northeast and subsiding more rapidly in that direction. Alternative explanations for the dramatic change in thickness from the Shubenacadie Basin (135 m, average, ranging 120 to 138 m) northwards across the TFZ (Fig. 13-2) to the Green Oaks area (670 m) have invoked differential subsidence of the Green Oaks area during sedimentation (Giles, 1977). The reconstruction of Figure 13-4A, on the other hand, accommodates this dramatic change in thickness over a much greater distance, parallel to the axis of the restored basins, and links satisfactorily with the Windsor Basin thickness (226 m) as an intermediate value.

In summary, Carboniferous stratigraphic relationships are better explained after reconstruction, and sedimentation patterns are compatible with the model. Thickness changes of major rock units, although equivocal, are consistent with deposition in an elongate trough, differentially subsiding towards the northeast.

PALEOMAGNETIC SUPPORT

Scotese et al. (1984) and Spariosu et al. (1984) argued that paleomagnetic declinations north and south of the Cobequid-Chedabucto Fault system were better aligned if southern Nova Scotia were rotated 20° to 25° in a clockwise direction to remove approximately 150 kilometres of post-Carboniferous dextral movement on that system. Data presented by these authors (Tables 13-1 and 13-2) can be used in a similar fashion to support the reconstruction of Figure 13-4A which invokes a counter-clockwise rotation of 13° of that part of southern Nova Scotia lying southeasterly of the Tobiatic Shear Zone.

Carboniferous paleomagnetic declinations for Shubenacadie Basin strata (Table 13-1) are closely comparable for both the Coldstream and Green Oaks Formations, and may be averaged to yield a value of 163.4° for the Tournaisian/Visean southeast of the TFZ.

Paleomagnetic data from the Cheverie Formation northwest of the TFZ reported by Scotese et al. (1984) and Spariosu et al. (1984) are in accord prior to correction for bedding tilt (Table 13-2). The corrected values of Spariosu et al., however, differ significantly from the two sites of Scotese et al., and are here accepted as the preferred values since more sites were considered (11) and their correction for tectonic tilt removing the effect of flexural-slip folding, more rigorous. The preferred mean value of palaeomagnetic declination (146°) is clearly divergent from the declination of 163.4° accepted for the Shubenacadie Basin (Fig. 13-1A). The reconstruction of Figure 13-4A implies counter-clockwise rotation on the broadly accurate TFZ satisfying major geological constraints and, at the same time, bringing these paleomagnetic declinations into much closer alignment.

This procedure in no way refutes the conclusion of Spariosu et al. (1984) that 20° to 25° of clockwise rotation of southern Nova Scotia along the Cobequid-Chedabucto system results similarly

Table 13-1. Paleomagnetic Data¹ - Coldstream Formation (Cheverie Equivalent) and Windsor Group, Shubenacadie Basin

FORMATION/GROUP	N/n	SITE IDENTIFIER	IN SITU		BEDDING DIRECTION	
			DECL.	INCL.	DECL.	INCL.
Coldstream	6/6	H3	154	29	--	--
Coldstream	6/6	H5	169	20	174	19
Coldstream	2/2	H14	162	12	160	19
Coldstream	6/6	H15	162	16	160	23
Coldstream Mean			161.8	19.3	164.7	20.3
Windsor	7/8	W1	163	08	164	12
Windsor	5/5	W2	169	05	168	11
Windsor	4/4	W4	165	03	--	--
Windsor	6/6	W6	161	18	162	16
Windsor	5/5	W8	173	17	172	14
Windsor	6/6	W9	165	21	163	15
Windsor	5/5	W10	164	19	162	13
Windsor	3/3	W11	154	27	149	20
Windsor Mean			164.3	14.8	162.9	14.4 ²
Shubenacadie Basin Mean (Tournaisian-Viséan)			163.5	16.3	163.4	16.2 ²

¹ Data from Scotese et al., 1984² Recommended values for Tournaisian/Viséan, Shubenacadie Basin

Table 13-2. Paleomagnetic Data - Cheverie Formation Windsor Basin

REFERENCE	N/n	SITE IDENTIFIER	IN SITU		BEDDING DIRECTION ¹		UNFOLDED ²	
			DECL.	INCL.	DECL.	INCL.	DECL.	INCL.
Scotese et al., 1984	1/1	C3	170	23	167	31		
Scotese et al., 1984	2/2	C4	158	11	156	17		
Spariosu et al., 1984	3/3	MCA	143.1	36.0			142	26
Spariosu et al., 1984	5/5	MCB	142.6	40.7			149	32
Spariosu et al., 1984	4/5	MCC	157.9	26.5			153	15
Spariosu et al., 1984	5/6	MCD ^{AF}	146.9	37.6			146	27
Spariosu et al., 1984	5/5	MCE	144.8	21.3			142	30
Spariosu et al., 1984	5/5	MCF	151.4	29.5			147	35
Spariosu et al., 1984	5/5	MCG	152.1	15.6			149	25
Spariosu et al., 1984	5/5	MCH	166.4	47.3			133	7
Spariosu et al., 1984	3/5	MCI	160.4	42.4			141	23
Spariosu et al., 1984	4/4	MCJ	180.6	56.9			139	22
Spariosu et al., 1984	6/6	MCK	165.1	29.5			162	27
Mean of Scotese et al., 1984			164	17	161.5	24	--	-- ³
Mean of Spariosu et al., 1984			155	35	--	--	146	25 ³

¹ Corrected for bedding tilt² Corrected using method of Ramsay (1967) to remove effects of flexural-slip folding³ Recommended values for Tournaisian, Windsor Basin

in a better alignment of Carboniferous paleomagnetic declination. It does, however, emphasize a major, as yet unexplained conflict in interpreted post-Visean relative movements on these two major fault systems which appear to merge at their eastern limits of onshore expression (Fig. 13-4A).

The rotation of paleomagnetic delineations of Early Carboniferous age offers support for the Devonian-Carboniferous reconstruction of Figure 13-4A. Comparison of similar data for Triassic rocks south and north of the Tobiatic Fault Zone (Laroche and Wanless, 1966; Carmichael and Palmer, 1968) shows no such rotation, with declinations in close agreement (Fig. 13-1A) when plotted on the present geological configuration of southern Nova Scotia. These data suggest no post-Triassic rotational movement on the TFZ and may therefore place an upper (younger) limit on major fault movement.

DISCUSSION

METALLOGENY

The similarity of Devonian-Carboniferous granitoid rocks of the Appalachians to Hercynian granitoids of western Europe has been emphasized by Strong (1974, 1980). Evidence is now available to show that the granitoid rocks of southern Nova Scotia have undergone petrological and metallogenic evolution closely comparable to that of Hercynian granitoids (Chatterjee and Muecke, 1982; Chatterjee et al., 1983) which host well known deposits of Sn, W and associated elements. The role of metallogenetically specialized leucogranites in this evolution is of fundamental importance in Hercynian Europe (Burnol, 1979; Fischendorf, 1978) as it is in Nova Scotia (Chatterjee et al., 1983).

Strong and Hanmer (1981) showed close spatial and genetic links between Carboniferous leucogranites and the South American Shear Zone of western Europe. Bell and Blenkinsop (1977) dated Carboniferous granites in Newfoundland which were either cataclastically deformed or intruded along zones of major weakness. Thus, Carboniferous granitoids elsewhere in the Appalachian-Hercynian fold belts relate in part to zones of Late Carboniferous deformation. Metallogenic domains of Chatterjee (1983) characterized by granophile mineralization (Fig. 13-6), with the single exception of the U-Cu-Ag domain parallel to the arcuate northwestern margin of the SMB, show a spatial link to the Tobiatic Fault Zone along a broad northeasterly zone stretching virtually the entire length of southern Nova Scotia.

Model lead ages for granophile deposits proximal to the TFZ range from 271 to 300 Ma (Fig. 13-6; Chatterjee, 1983), entirely consistent with post-Wedgeport Pluton (317 Ma) sinistral movement on the zone. The genetic links with specialized leucomonzogranites, age of mineralization and proximity to the Tobiatic Fault Zone, suggest that the latter might provide a mechanism to localize the mineralizing fluids and perhaps the parent leucomonzogranites applying the model of Strong and Hanmer (1981)

for the origin of leucogranites through faulting, frictional heating, fluid loss and fractional melting. In support of this suggestion, metallogenic domains extending to the northeast and southwest well beyond the limits of the major granitoid bodies cut by the TFZ nevertheless exhibit granophile metal associations and appear to relate to specialized granites at depth. In counter argument, however, leucogranites described by Strong and Hanmer are not entirely comparable to the leucomonzogranites of southern Nova Scotia, and many of the leucomonzogranites are spatially unrelated to the TFZ or known parallel systems (see Chatterjee, 1983). An alternative explanation of the observed relationships is the control of migration of both magmatic and hydrothermal fluids by the TFZ as a major zone of weakness, without invoking granite origin through the shear mechanism itself. An emerging recognition of more widespread shear deformation of granitoid rocks of the SMB may allow this interpretation to be applied on a broader scale to explain the distribution of leucomonzogranites unrelated to the Tobiatic Fault Zone.

That the Tobiatic Fault Zone has served as a preferred pathway for the migration of metal-bearing fluids is confirmed by the widespread occurrence of Ba-Pb-Zn mineralization in Carboniferous strata cut by the TFZ along the northern margin of the Shubenacadie Basin (Fig. 13-6). These deposits occur as replacement bodies and open-space filling of fracture systems reflecting movement on the TFZ, and are seemingly unrelated to granitoid rocks at depth. Thus the significance of the zone extends beyond granite-related models for metallogenesis.

CONSTRAINTS ON PALEOMAGNETIC/TECTONIC MODELS

Apparent discrepancies between Carboniferous paleomagnetic results from Maritime Canada ("Acadia") and cratonic North America have led to the suggestion of 1000 to 1500 kilometres of sinistral displacement of Acadia with respect to the craton during the Carboniferous (Kent and Opdyke, 1979; Irving, 1979; van der Voo et al., 1979; Kent, 1982; Scotese et al., 1984; and Spariosu et al., 1984). Irving and Strong (1984) have recently shown that no measurable discrepancy exists, based on new determinations on cratonic Visean strata of western Newfoundland, and have therefore argued that such major displacement of Acadia could not have occurred during the Visean or later. The apparent discrepancy was attributed by Irving and Strong to widespread Kiaman overprinting of the cratonic paleomagnetic record.

The reconstruction of Figure 13-4A offers additional constraints for the model of Kent and Opdyke (1979) in that it suggests no major sinistral fault movement in southern Nova Scotia from the time of emplacement of the S-M-L complex (372-360 Ma) at least to Namurian time. Indeed, rapid unroofing of the SMB prior to Horton Group sedimentation (Chatterjee and Muecke, 1982) suggests that vertical movements were regionally more important. Accepting a figure of 325 Ma as the base of the Namurian (George et al., 1976) this 35 million year period is not compatible

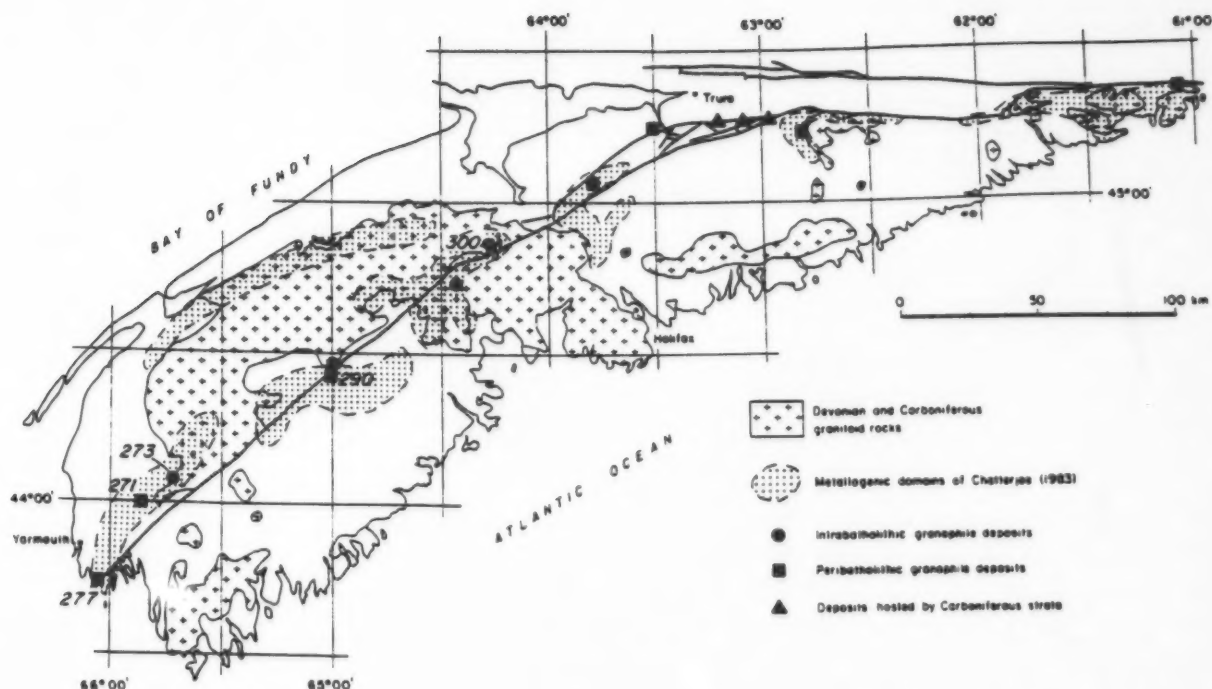


Figure 13-6: Polymetallic metallogenic domains of granophis character of Chatterjee (1983) illustrating their general proximity to the Tobiatic Fault Zone. Note the occurrence of selected peribatholithic, intrabatholithic and other mineral deposits within and proximal to the zone. Numbers are model lead ages of Chatterjee (1983) believed to date mineralizing events related to fault movements.

with major regional stress systems which could result in sinistral displacement of Acadia relative to cratonic North America during the Late Devonian or earliest Carboniferous. If Irving and Strong (1984) are correct in denying the possibility of Visean or post-Visean displacement, then the model of Kent and Opdyke indeed appears to have little temporal scope in southern Nova Scotia. This conclusion cannot, however, refute possible sinistral displacement of large magnitude on other major fault systems closer to cratonic North America during Late Devonian to earliest Carboniferous time. The 110 kilometres of sinistral movement on the Tobiatic Fault Zone postulated for post Visean time, which appears initially to offer distinct possibilities as part of the cumulative displacement of Acadia relative to the Craton, must now be considered as an expression of Hercynian tectonism pertinent mainly to southern Nova Scotia, and as yet, unassimilated within the regional tectonic framework.

Sinistral displacement of post-Visean (?Westphalian) age on the Tobiatic Fault Zone is an apparent tectonic anomaly in a region dominated by dextral fault movement during the Late Devonian and Carboniferous (Webb, 1963, 1969; Keppie, 1977, 1982). Webb's (1963) careful analysis of the Harvey-Hopewell Fault (HHF) in southern New Brunswick confirms that the TFZ is not unique in its sinistral movement history. Webb documented 16 kilometres of post-Visean

sinistral offset on the HHF subsequent to an earlier Carboniferous dextral offset of 80 kilometres on the same zone, and stated that such divergent motions should be expected more commonly on parallel adjacent northeasterly trending faults.

In Atlantic Canada, these complex movement histories are overshadowed by by predominant dextral shear on most of the major faults (Webb, 1969) including the Cobequid-Chedabucto Fault System (Eisbacher, 1969, 1970) which appears ubiquitously on tectonic reconstructions of the Appalachian-Hercynian fold belt. Resolution of these contrasting movement histories must be addressed as a local problem without undue emphasis on plate tectonic implications. Such resolution, however, will be extremely important in understanding the role of tectonics in the metallogeny of Nova Scotia, particularly in light of the clear relationship of the TFZ to mineralizing events.

ADDITIONAL FOOD FOR THOUGHT

The reconstruction of Figure 13-4A offers an entirely new perspective on granitoid rocks comprising the South Mountain and Musquodoboit Batholiths and the Liscomb Pluton. The very large elliptical caldron (S-M-L complex; Pitcher, 1978) suggested by the reconstruction affords possible new insight into mechanisms of granite emplacement which can be tested against gravity

data. The westerly deflection of Bouguer Gravity anomaly in the New Ross area (Fig. 13-18) and the separation of a small discrete Bouguer Gravity low west of New Ross as a consequence of the reconstruction (Fig. 13-48), raises the possibility that the root zone of the S-M-L complex might be expressed by that low. Detailed analysis of existing geophysical data may test this possibility.

The emplacement of large granite batholiths along major crustal fault zones is well documented (Pitcher and Bussell, 1977; Pitcher, 1978). Keppie and Chatterjee (1982) noted the apparent absence of any such zone related to the SMB. Critical assessment of the TFZ as a possible re-activated pre-Devonian crustal fault zone, similar in temporal scope and nature to the Cobequid-Chedabucto Fault System, could lead to models for emplacement of the S-M-L complex along a major crustal fault.

Carboniferous palaeogeographic reconstructions can and should be tested more rigorously against the model of Figure 13-4A. The development of an ancestral Shubenacadie (and Musquodoboit) Basin in the core of the S-M-L complex can be modeled using caldron subsidence as a contributing factor.

Present conflicts in structural interpretation should be addressed, in particular for Late Carboniferous-Permian time, in an attempt to resolve presently incompatible fault movements though critical re-examination of existing and new data. Time of fault movement must be more rigorously determined. Towards this end, possible sinistral offset of metallogenic domains with dated mineralizing events should be critically assessed.

As with most investigations, recognition of the Tobistic Fault Zone and the reconstruction of Figure 13-4A appears to have generated as many questions as answers and will hopefully stimulate others to address these important aspects of Nova Scotia geology.

SUMMARY AND CONCLUSIONS

The Tobistic Fault Zone is defined as a major zone of sinistral shear extending from the eastern extremities of southern Nova Scotia, westerly and southwesterly to the Yarmouth area, a distance in excess of 400 kilometres. Possible extension into the Georges Basin Fault System and parallel faults in the Gulf of Maine, adds a further 350 kilometres to the length of the shear zone.

Geological evidence suggests approximately 110 kilometres of sinistral offset in post-Visean pre-Triassic time. Removal of this displacement results in a Devonian/Carboniferous reconstruction which resolves the enigmatic "banana-shape" of the SMB by making it a part of a much larger, elliptical plutonic complex including the Musquodoboit Batholith, Liscomb Pluton and several small, unnamed plutons. The reconstruction retains close spatial relationships of Carboniferous facies belts by inferring the longitudinal extension of the Shubenacadie Basin northeasterly into the Windsor Basin.

The Devonian-Carboniferous reconstruction merges regional anomalies in Bouguer Gravity and air-borne radiometrics and is compatible with available paleomagnetic data. Metallogenic domains characterized by granophile mineral deposits including Sn, W, Cu, Pb, Zn, Ag, Au, Sb and U, are preferentially distributed along the trace of the shear zone. Mineralizing events dated by model lead ages (270 to 300 Ma) are compatible with post-Visean pre-Triassic shear movements and suggest a genetic link to the fault zone.

Peribatholithic granophile deposits along the trace of the fault zone are genetically linked to specialized granitoids which may have been preferentially emplaced in the zone, or in parallel zones of shear. Deposits of Ba-Pb-Zn in Carboniferous strata show a similar spatial link to the TFZ but may owe their origin to processes unrelated to plutonic activity.

Postulated sinistral movements on the Tobistic Fault Zone are too young to support paleomagnetic/tectonic models for major sinistral displacement of Maritime Canada relative to cratonic North America, in the face of recent arguments that such movement could not have taken place later than Visean time.

The timing and sense of movement on the Tobistic Fault Zone is not yet reconciled with opposing movements of generally similar age on adjacent or shared fault systems (or both). Although the reconstruction of this study appears to rationally explain many geological and geophysical features of southern Nova Scotia, it simultaneously creates many opportunities for future, more comprehensive research dealing with granitoid rocks, mineral deposits, Carboniferous stratigraphy, sedimentology and paleogeography, as well as regional structural geology and tectonics.

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